Bulletin
of
The Geological Society of America

VOLUME 63 July 1952 NUMBER 7

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Subscription $15.00 per year.

Publication Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publication should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117 Street, New York, N. Y.

NOTICE—Four weeks notice of change of address is necessary. In accordance with the rules established by the Council, claims for for non-receipt of the preceding number of the Bulletin must be sent the Secretary of the Society within three months of the date of the receipt of this number in order to be filled gratis.

Entered as second-class matter at the Post-Office at Baltimore, Md., under the Act of Congress of July 16, 1894.

Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 8, 1918.

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 63, PP. 649-688, 6 PLS., 3 FIGS. JULY 1952

PETROGRAPHIC RELATIONS IN SOME TYPICAL BAUXITE AND DIASPORE DEPOSITS

BY VICTOR T. ALLEN

ABSTRACT

Three processes that operate in the formation of bauxite and diaspore deposits are (1) desilication, in which silica is removed from clays, from feldspars, and from other aluminous minerals; (2) migration, in which clay minerals and hydrous aluminum oxides are transported within the deposit from the place where they were formed to their present position; (3) resilication, in which silica is added to gibbsite to form a hydrous aluminum silicate or clay.

The desilication of feldspars and parent rocks is usually a two-stage process in which clay minerals such as kaolinite, halloysite, and nontronite are formed first and then silica is removed from them to form gibbsite, boehmite, and diaspore. In the desilication of sedimentary clays, the original structure and texture of the clay are preserved. The formation of bauxite directly from feldspars and nepheline, which has been reported at some localities, does not appear to be the dominant process at the deposits studied during this investigation. Replacement of the clay minerals by gibbsite, boehmite, and diaspore is gradual and involves decrease in silica and increase in alumina, titania, and porosity. Transitional zones of clay-retaining structures of the parent rocks indicate that clay was an intermediate stage in the formation of many bauxite deposits. Desilication takes place in a warm moist climate through the agency of carbonic acid, alkaline carbonates, and magnesium bicarbonate solutions, either as surface weathering or by ground water diffusing through masses of clay of varying permeability and porosity.

Clay minerals and hydrous aluminum oxides migrate along cracks and open spaces in the clays and bauxite. Pipes and nodules of gibbsite lacking the structure of the parent rock are formed by migration, concentration, and crystallization of secondary gibbsite.

The occurrence of cellular kaolin with a structure interpreted as inherited from gibbsite is considered evidence of resilication of gibbsite to kaolin in the bauxite district of Georgia. Dense opaque white to gray kaolinite and some of the "chimney rock" of Georgia may have formed by resilication, which takes place under conditions of poor drainage in low-lying areas.

The diaspore of Missouri, Pennsylvania, and Washington and the boehmite of Missouri, Kentucky, Washington, and California were formed from flint clays composed of finely divided kaolinite having low plasticity and breaking with a conchoidal fracture. The gibbsite of the bauxite deposits of Georgia and Alabama was formed from a kaolinitic clay of the hard or semihard type. The chief difference between the flint clays and the hard types of Georgia and Alabama is that the very fine particles of flint clay lack the degrees of symmetry in the stacking of the molecular sheets of kaolinite. The low plasticity of flint clays permitted the retention of openings and cracks without collapse, and this aided their alteration. Possible controlling conditions for the formation of gibbsite, boehmite, and diaspore include temperature, nature of the chemical systems, time, pressure, and drainage.

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Introduction

A project grant from the Penrose Bequest of The Geological Society of America in 1941 enabled the writer to collect samples of clays with and without associated bauxite minerals from Arkansas, Alabama, Georgia, Florida, North Carolina, South Carolina, and New Jersey and to study their petrographic relations. Previously a petrographic investigatigation of the flint and diaspore clays of Missouri (Allen, 1935a) had revealed that the monohydrate of aluminum, diaspore, was the chief high-alumina mineral formed from the kaolinitic flint clays of Missouri. The only occurrence of gibbsite in the diaspore district of Missouri (Allen, 1935b) was in a soft oölitic clay that is not a typical flint clay and that contains too little gibbsite to be recorded in an x-ray diffraction pattern. The published descriptions of the bauxite deposits of the United States, which are associated with sedimentary kaolins, listed the chief high-alumina mineral in these deposits as the aluminum trihydrate, gibbsite. This provoked the question of why disapore was formed from kaolinitic clays in Missouri and gibbsite from the kaolinitic clays of Georgia and other States.

Petrographic and x-ray studies of the samples collected in 1941 indicated that boehmite, the monohydrate of aluminum first observed in bauxite by Böhm (1925) and named by De Lapparent (1927), was present at one of these localities and was reported from another (Böhm, 1925). McQueen (1943) described boehmite in the newly discovered flint clay deposits near Swiss and Stolpe, Gasconade County, Missouri. Thus, the study of the occurrence and relations of boehmite with the same chemical composition as diaspore but with a different x-ray pattern was undertaken as a part of this project.

Specimens of weathered granite and syenite from North Carolina, Georgia, and Arkansas were also collected and examined to determine the minerals formed during the initial stages of weathering of feldspars and other minerals of the parent igneous rocks. With the announcement by Goldman and Tracey (1946) that "Much evidence is accumulating to support the belief that at least under certain conditions gibbsite precedes kaolin in the weathering of all but the most silicic igneous rocks," the impor-

tance of this phase of the study was greatly increased.

From 1942 to 1946, as Commodity Geologist in charge of high-alumina clay investigations of the U. S. Geological Survey, the writer visited many of the bauxite and high-alumina clay deposits of the United States. Some new occurrences of gibbsite and boehmite as well as progress have been reported (Allen, 1945; 1946b; 1947). Work has continued on the project. This paper summarizes the results of the writer's investigations.

Acknowledgments

Field collecting of specimens during the summer of 1941, the making of 60 x-ray patterns by Professor John W. Gruner of the University of Minnesota, and the completion of four chemical analyses were financed by a project grant from the Penrose Bequest of The Geological Society of America in 1941. The U. S. Geological Survey has generously permitted the inclusion of information collected by the writer when he was in charge of high-alumina clay investigations of the U. S. Geological Survey from 1942–1946.

Chemical analyses of Georgia bauxite samples by L. H. Turner were provided through the generosity of Captain Garland Peyton, Director of the Georgia Division of Mines, Mining and Geology. The guidance and companionship of A. S. Furcron made the field collecting of 1941 in Georgia pleasant and profitable. Chemical analyses of Alabama bauxite samples by R. S. Hodges were completed through the kind arrangements of Stewart J. Lloyd, Assistant State Geologist of Alabama. The co-operation of N. L. Shepard, R. C. Cross, A. D. Wilson, and other staff members of the Aluminum Company of America and subsidiaries of that company; F. E. Smith and Richard Janatka of the Harris Clay Company, and Kaolin, Incorporated, of North Carolina; Charles J. Vitaliano, then of Rutgers University; and several pit and mine operators, who aided in the collection of bauxite and clay samples, is gratefully acknolwedged.

X-ray patterns by J. M. Axelrod and differential-thermal curves by G. T. Faust made in the laboratories of the U. S. Geological Survey were helpful. Frequent discussions, many suggestions, and a critical reading of the manuscript were graciously contributed by Dr. Clarence S. Ross and Dr. George T. Faust of the U. S. Geological Survey.

Purpose and Scope of the Paper

The purpose of this paper is to record the minerals identified in bauxite and high-alumina clays by the petrographic, x-ray, chemical, and differential-thermal methods utilized during this investigation. The relation of the minerals to each other in the deposits is discussed, and a mode of origin compatible with the observed relations is proposed.

During this investigation, three fundamental processes concerned with the formation of bauxite and high-alumina clay have been recognized. These processes are discussed in the order of their relative importance under the following headings:

(1) Desilication is the removal of silica from feldspars, clays, or other minerals to form new clay minerals or hydrous aluminum oxides.

(2) Migration or movement of clay minerals and hydrous aluminum oxides involves their transportation within the deposit from the place where they were formed to their present position.

(3) Resilication is the union of silica with gibbsite or some hydrous aluminum oxide to form a hydrous aluminum silicate.

Desilication of Parent Rocks and Clays

Spruce Pine District, North Carolina

The Spruce Pine district (Fig. 1, loc. 1) includes parts of Avery, Mitchell, and Yancey counties, North Carolina.[1] The residual clay of this area has been derived by the weathering of a coarse-grained granite, called alaskite (Hunter, 1940), and the associated pegmatites. The weathering belongs to an early incomplete erosion cycle rather than to the present cycle, and its products are found on a well-developed strath or old erosion level along the North Toe River and its main tributaries. The age of the oldest erosion level is uncertain, but Hunter (1940, p. 101) states that Edwin C. Eckel con-

[1] Places not shown on the index map (Fig. 1) may be found on maps of the U. S. Geological Survey, the U. S. Army Engineers, and the U. S. Tennessee Valley Authority.

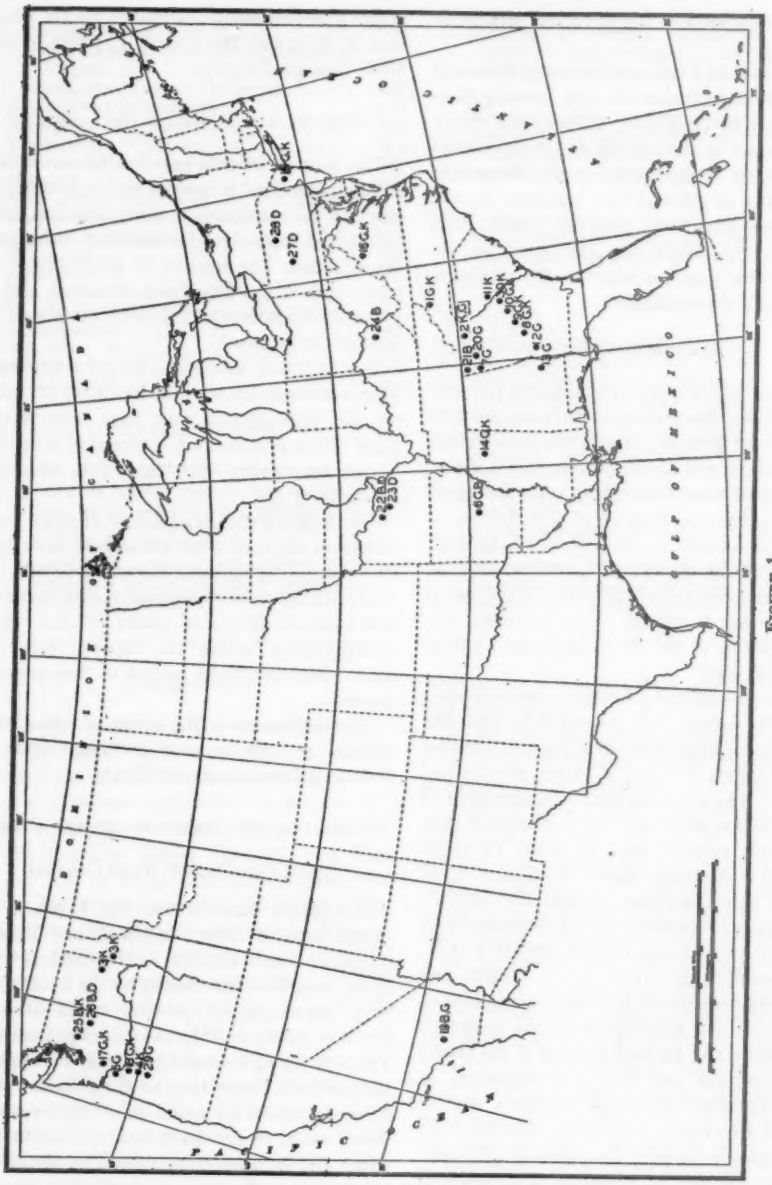
GULF OF MEXICO
PACIFIC OCEAN
28D
27D
18GK
1GK
11K
24B
14GK
6GB
23B
23D
25BK
26BD
17GK
18BG

FIGURE 1

siders the oldest terrace level Tertiary. This concept is shared by Parker (1946), who mapped the kaolin deposit as a part of the high-alumina clay program of the U. S. Geological Survey.

Weathering of the original minerals of the granite has been selective. The structural control afforded by increased permeability, such as exists along fractured quartz veins, allowed ground-water solutions to decompose minerals at depths greater than those in less favorable parts of the deposit. Locally, decomposition extends downward 100 feet or more, but the commercial deposits of kaolin are limited to a depth of about 40 or 50 feet, because an appreciable quantity of partly decomposed rock appears at that depth. Orthoclase, which is sparsely represented, decomposes to kaolinite most easily. Microcline, albite, muscovite, and quartz resist decomposition and remain unaltered or only slightly altered even when the surrounding minerals are completely changed to kaolin. Biotite alters to hydromica, and garnet to brownish spots containing iron or manganese oxides.

Several investigators (Somers, 1921; Ries, 1927) assume an isomorphous gradation between kaolinite and muscovite which is marked by a gradual loss of potash and addition of water, whereby in the process of weathering hydromica may represent a transitional stage toward kaolinite. Hickling (1909) studied the China clay of Cornwall, England, where kaolinite and mica make up 90 per cent of the mass and the relative amounts of kaolinite, hydromica, and muscovite are difficult to estimate. Emphasis is placed on the observation of prisms composed of mica at one end and kaolinite at the other, and on the experiment of Johnstone (1889), who found that muscovite changed to hydromuscovite after immersion for 12 months in water saturated with carbon dioxide. Galpin (1912) reports "ribs" of higher index and birefringence intergrown with those of kaolinite and showing practically every gradation between kaolinite and muscovite. Ross and Kerr (1931) found a mineral with moderately high index and birefringence intergrown with kaolinite and considered it was similar to kaolinite except that it contained one molecule of water instead of two. Ample evidence in recent papers demonstrates that the clay minerals initially formed from a specific mineral or rock may differ with

FIGURE 1.—INDEX MAP OF THE UNITED STATES SHOWING LOCALITIES WHERE RESIDUAL AND SEDIMENTARY KAOLIN DEPOSITS (K), GIBBSITE (G), BOEHMITE (B), AND DIASPORE (D) DEPOSITS WERE STUDIED

(1) Spruce Pine district, North Carolina
(2) Rabun County, Georgia
(3) Excelsior, Wash.; Latah County, Idaho
(4) Mehama district, Marion County, Oregon
(5) Ferruginous bauxite in Washington, Columbia and Multnomah counties, Oregon
(6) Bauxite district, Arkansas
(7) Bauxite district, Floyd County, Georgia
(8) Bauxite and kaolin district, Twiggs County, Georgia
(9) Kaolin district, Wilkinson County, Georgia
(10) Kaolin district, Washington County, Georgia
(11) Kaolin district, Aikin district, South Carolina
(12) Andersonville bauxite district, Macon and Sumter counties, Georgia
(13) Bauxite district, Barbour and Henry counties, Alabama
(14) Bauxite district, Pontotoc and Union counties, Mississippi
(15) Kaolin with gibbsite district, Woodbridge, New Jersey
(16) Bauxite district, Augusta County, Virginia
(17) Castle Rock deposit, Cowlitz County, Washington
(18) Molalla deposit, Clackamas County, Oregon
(19) Kaolin deposit, Riverside County, California
(20) Bartow County clay district, Georgia
(21) Walker County clay district, Georgia
(22) Gasconade County district, Missouri
(23) Diaspore district, Phelps, Maries, Crawford counties, Missouri
(24) Olive Hill district, Carter County, Kentucky
(25) Durham district, King County, Washington
(26) Cle Elum deposit, Kittitas County, Washington
(27) Curwensville district, Clearfield County, Pennsylvania
(28) Clinton County clay district, Pennsylvania
(29) Gibbsite deposit, Salem, Oregon
(30) Glascock County clay deposit, Georgia

the conditions and the reacting solutions under which the alteration takes place.

To test Hickling's idea that feldspars weather first to mica and then to kaolinite, numerous specimens of the residual clay and the associated weathered feldspars were examined in oils of known indices of refraction by the immersion method. The most satisfactory materials for this test were selected, and thin sections were made of them and of representative clays. Examination of these thin sections showed orthoclase altered to kaolinite without a suggestion of the formation of intermediate products. Microcline, albite, and parts of perthitic intergrowths are fresh or slightly altered in the same sections where orthoclase and plagioclase containing substantial amounts of the anorthite molecule were completely changed to kaolinite (Pl. 1, fig. 1). The alteration product that locally clouds the surface of the feldspars has low interference colors, which are characteristic of kaolinite rather than of hydromica. Muscovite, which is a primary constituent of the pegmatitic phases of the granite, resists weathering, and unaltered or only slightly altered muscovite is recovered as a by-product during the washing and preparation of the kaolin for utilization. The washed kaolin approaches closely the composition of kaolinite (Table 1) and contains only a small proportion of fine quartz and muscovite.

Two specimens of the clay from Spruce Pine, North Carolina, were submitted to Professor John W. Gruner to determine whether x-ray methods would reveal relict hydromicas too fine to be observed by optical examination. The x-ray patterns indicated that one clay is composed of halloysite and kaolinite, and the other is a mixture of kaolinite, gibbsite, and microcline. The gibbsite recorded in the x-ray pattern is too fine-grained to be visible in thin section. This suggests that finely divided gibbsite is formed by the weathering of the kaolinite, which clouds irregular patches of the microcline perthite. Since finely divided gibbsite should respond first to resilication, it would not be expected to survive when other gibbsite in the specimen had changed to kaolinite or halloysite. The x-ray patterns and petrographic evidence are in agreement that halloysite and kaolinite are the first minerals formed during the initial stages of weathering of the feldspars in these North Carolina deposits. This conclusion agrees with earlier observations of the writer on the feldspars in the granitic rocks of the Sierra Nevada of California (Allen, 1929).

A "hydromica" from Waynesville, North

TABLE 1.—CHEMICAL COMPOSITION OF WASHED KAOLIN, HYDROMICA, AND A GIBBSITE-KAOLINITE-FELDSPAR SPECIMEN

	1	2	3	4	5	6
SiO_2	47.94	46.18	46.5	41.98	40.79	21.65
Al_2O_3	37.02	38.38	39.5	29.54	29.98	49.65
Fe_2O_3	0.60	0.57		9.81	8.07	0.85
FeO	—	—		0.67	2.48	—
TiO_2	0.02	0.04		1.20	1.28	none
CaO	0.30	0.37		none	0.45	none
MgO	0.07	0.42		1.77	2.71	none
Na_2O	0.06	0.10		0.36	0.38	0.21
K_2O	1.25	0.58		2.06	3.47	3.33
Others		0.08				
H_2O − 110°C	13.03	13.28	14.0	1.12	1.20	0.40
H_2O + 110°C				11.32	9.34	23.73
Total	100.29	100.00		99.83	100.15	99.82

1, 2. Washed kaolin, Spruce Pine, North Carolina. Analyses furnished by Harris Clay Company and Kaolin, Inc. (Hunter, 1940).

3. Theoretical composition of kaolinite to satisfy formula $Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$.

4. Hydromica with a little kaolinite. Alteration product of biotite, Lunday mine near Micaville, North Carolina. Analyst, Joseph J. Fahey.

5. Hydromica, Herren property, Waynesville, North Carolina. Analyst, George Steiger (Bayley, 1925).

6. Gibbsite-kaolinite-feldspar specimen derived by weathering of feldspar. W. E. Norton property, Land Lot 20, District 3, Rabun County, Georgia. Analyst, Joseph J. Fahey.

Carolina (Bayley, 1925), appears to have resulted from the alteration of muscovite. The chemical composition of this specimen determined by George Steiger (Table 1, No. 5) is similar to that of a "Hydromica" collected by the writer from the Lunday mine, near Micaville, North Carolina, and analyzed by Joseph J. Fahey (Table 1, No. 4). The hydromica from the Lunday mine was derived by the weathering of biotite, as shown by some of the less altered sheets of biotite, which retain their brownish color near the center and grade outward into bronze to cream-colored hydromica. Sheets uniformly colored bronze were selected by the writer for chemical analysis and for x-ray examination by Professor John W. Gruner, who reported they are composed of hydromica with a little kaolinite. Since the hydromica was embedded in white kaolinite derived from the weathering of feldspars, some kaolinite might have been included between sheets, but none could be detected. It appears that some of this hydromica derived from biotite has changed to kaolinite. Anauxite, which is isomorphous with kaolinite, has been described by the writer as an alteration product of biotite in California (1928).

The optical properties of the hydromica from the Lunday mine are: $\alpha = 1.572$; $\gamma = 1.618$; $\gamma - \alpha = 0.046$; 2V approximately 0°; optical character, negative; some sheets contain spots of yellow iron oxides. The optical properties of the hydromica from Waynesville were not recorded by Bayley, but, because its iron, potassium, and magnesia are higher and its water content is lower, one would expect its indices of refraction to be slightly higher than those determined by the writer for the hydromica from the Lunday mine. Furthermore, since the indices of refraction of the hydromica from the Lunday mine are higher than those of any muscovite listed by Larsen and Berman (1934), it is unlikely that the hydromica from Waynesville could have resulted from the alteration of muscovite, because decrease of potassium and iron and increase of water would decrease rather than increase the indices of refraction of the resulting hydromica. For this reason and because muscovite remains stable and even in tiny sheets retains its characteristic indices of refraction and large optic angle, the writer believes that the hydromica from Waynesville was derived from biotite rather than muscovite.

Rabun County, Georgia

Specimens of microcline-perthite pegmatite from a trench 19 to 22 feet below the surface at the W. E. Norton property, Land Lot 20, District 3, near Clayton, Rabun County, Georgia, were furnished for this investigation by W. H. Yoho (Fig. 1, loc. 2). Most of these consisted of fresh pink microcline-perthite, but on some specimens tiny veinlets and pockets of white powder were surrounded by fresh feldspar. Some of the powder was entirely halloysite or kaolinite; other parts were a mixture of these minerals with gibbsite or feldspar. Two of these powders were submitted to Professor John W. Gruner, who reported that x-ray patterns indicated one was about half halloysite and half gibbsite and the other was crystalline gibbsite with a little microcline and kaolinite. The latter specimen consisted of spherical groups of radiating crystalline gibbsite up to an eighth of an inch in diameter surrounded by fine-grained kaolinite that formed small patches in the microcline-perthite. Several of the spherical masses were isolated and submitted to Joseph J. Fahey for a chemical analysis.

Calculation of this analysis (Table 1, No. 6) indicates 60.5 per cent of gibbsite. If the K_2O and Na_2O in this analysis is combined with enough Al_2O_3 and SiO_2 to form microcline (K_2O, Na_2O, $Al_2O_3 \cdot 6SiO_2$), the sample contains 20.91 per cent microcline. If the remaining silica (8.10 per cent) is combined with sufficient water and alumina to form kaolinite ($Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$), the sample would contain 17.56 per cent kaolinite. The remaining Al_2O_3 (38.96 per cent) and the remaining H_2O (21.54 per cent) would combine to form 60.5 per cent gibbsite ($Al_2O_3 \cdot 3H_2O$). If these last figures are recalculated on the basis of 100 per cent, the results approach closely the 65.4 per cent Al_2O_3 and the 34.6 per cent H_2O required by the formula of gibbsite. X-ray, optical, and chemical data agree that this specimen is a mixture of gibbsite, kaolinite, and microcline. Petrographic observations indicate that kaolinite and halloysite are formed as the first weathering product of the feldspar. In some of these specimens, the change ended with the formation of these kaolin minerals; in others, gibbsite was formed from the kaolin minerals by desilication. The radiating groups of crystalline

gibbsite have sharp contacts with the surrounding fine-grained kaolinite and were developed by growing outward at the expense of the kaolinite. This process probably took place during an old incomplete erosion cycle on the rocks of the Piedmont plateau rather than during the recent cycle, but the exact age of the weathering is unknown. Most of the weathered feldspathic rocks of the Piedmont plateau examined by the writer contain only clay minerals, which dot the feldspars and extend as a network across them.

Columbia River Region of Oregon, Washington, and Idaho

In the Columbia River region of Oregon, Washington, and Idaho (Fig. 1, loc. 3), plagioclase of basaltic lavas is altered to kaolinite (Pl. 1, fig. 2), halloysite, endellite (Pl. 1, fig. 3), and nontronite. However, nontronite is also formed from augite, iddingsite, basaltic glass, and palagonite under conditions of poor drainage and in the presence of magnesium. The kaolin minerals, kaolinite and halloysite, are formed under good drainage and neutral or slightly acid conditions. Halloysite and kaolinite occur in the same vertical section, which suggests they were formed in this region under nearly similar conditions of weathering. In some specimens, plagioclase is altered along cleavages and cracks to clay minerals that extend as irregular zones across the crystals (Pl. 1, figs. 2, 3). All stages from partly altered to completely kaolinized plagioclase have been

PLATE 1.—PHOTOMICROGRAPHS OF ALTERED FELDSPARS

FIGURE 1. ORTHOCLASE AND PLAGIOCLASE COMPLETELY ALTERED TO KAOLINITE (K, DARK). MICROCLINE (M, WHITE) UNALTERED

Kaolin Inc. mine, 5 miles S. E., Spruce Pine, N. C.

FIGURE 2. EARLY STAGE OF ALTERATION OF PLAGIOCLASE TO KAOLINITE (K)

(K) cuts plagioclase laths as veinlets and embayed areas leaving relict areas of plagioclase (white) in a weathered basalt. Nontronite (N) formed from basaltic glass, Ilmenite (I). Drill hole 19, depth 58 feet, Excelsior clay deposit, Washington.

FIGURE 3. EARLY STAGE OF ALTERATION OF PLAGIOCLASE (P) TO ENDELLITE ALONG CLEAVAGES AND VEINS

Augite, iddingsite, and ilmenite are unaltered. Weathered basalt, 8 miles south of Salem, Oregon.

FIGURE 4. LATHS OF FELDSPAR (F) PARTLY ALTERED TO AND CUT BY VEINLETS OF HALLOYSITE

Areas of halloysite (H) fill space between feldspars. Altered syenite. Bauxite district, Arkansas.

FIGURE 5. EARLY STAGE OF WEATHERING OF AN AEGIRITE SYENITE

Feldspar (F) is partly altered to halloysite white, but aegirite retains original optical properties. Bauxite district, Arkansas.

FIGURE 6. LATH OF FELDSPAR ALTERED TO KAOLINITE (K) AND GIBBSITE (DARK G) PARALLEL TO THE CLEAVAGE

Center of boulder surrounded by bauxite. Neilson underground mine, Bauxite district, Arkansas.

PLATE 2.—PHOTOMICROGRAPHS OF SEDIMENTARY CLAYS

FIGURE 1. PISOLITE IN WHICH KAOLINITE PLATES SHOW "WORM" STRUCTURE

Soft kaolin. Paragon Mines, Langley, South Carolina.

FIGURE 2. "PISOLITE" OF CLAY REWORKED FROM AN UNDERLYING FORMATION

Quartz grains are more abundant near the margin that those in the matrix or near the center of the "pisolite". At top small veins of gibbsite (G) cut the edges of the "pisolite". Drill core of hole 29, Pinedale, Mississippi.

FIGURE 3. PISOLITE FORMED BY CONTRACTION CRACKS AS CLAY DRIES

Open spaces are developed (O, white) and gibbsite forms in and near them. Andersonville bauxite district, Georgia.

FIGURE 4. PISOLITES FORMED BY CONTRACTION IN BAUXITE WITH 67 PER CENT Al_2O_3

Gibbsite is concentrated along the cracks. Depth 52 feet. Andersonville bauxite district, Georgia.

FIGURE 5. GRAINS OF VOLCANIC ROCKS ALTERED TO KAOLINITE (K); GIBBSITE (G) FORMED ALONG EDGES OF GRAINS

Castle Rock, Washington.

FIGURE 6. GRAINS OF VOLCANIC ROCKS ALTERED TO NONTRONITE (N); GIBBSITE (G) FORMED ALONG EDGES OF GRAINS

Castle Rock, Washington.

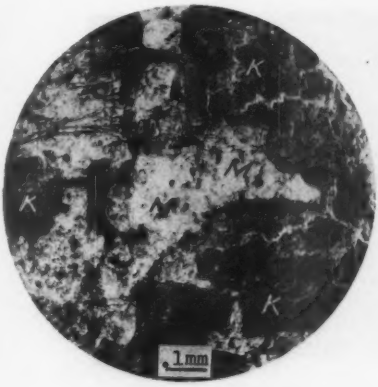

FIGURE 1

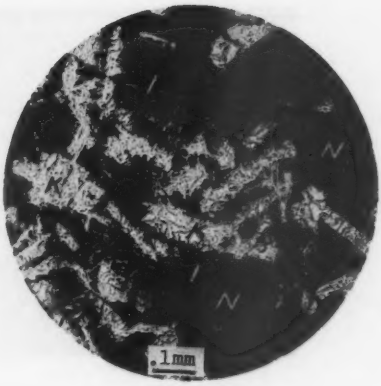

FIGURE 2

FIGURE 3

FIGURE 4

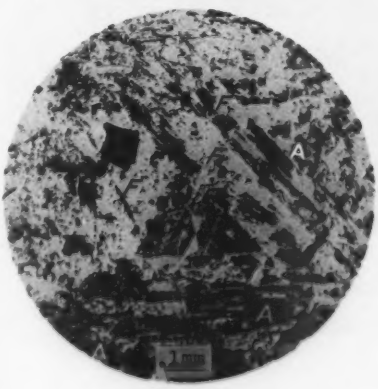

FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS OF ALTERED FELDSPARS

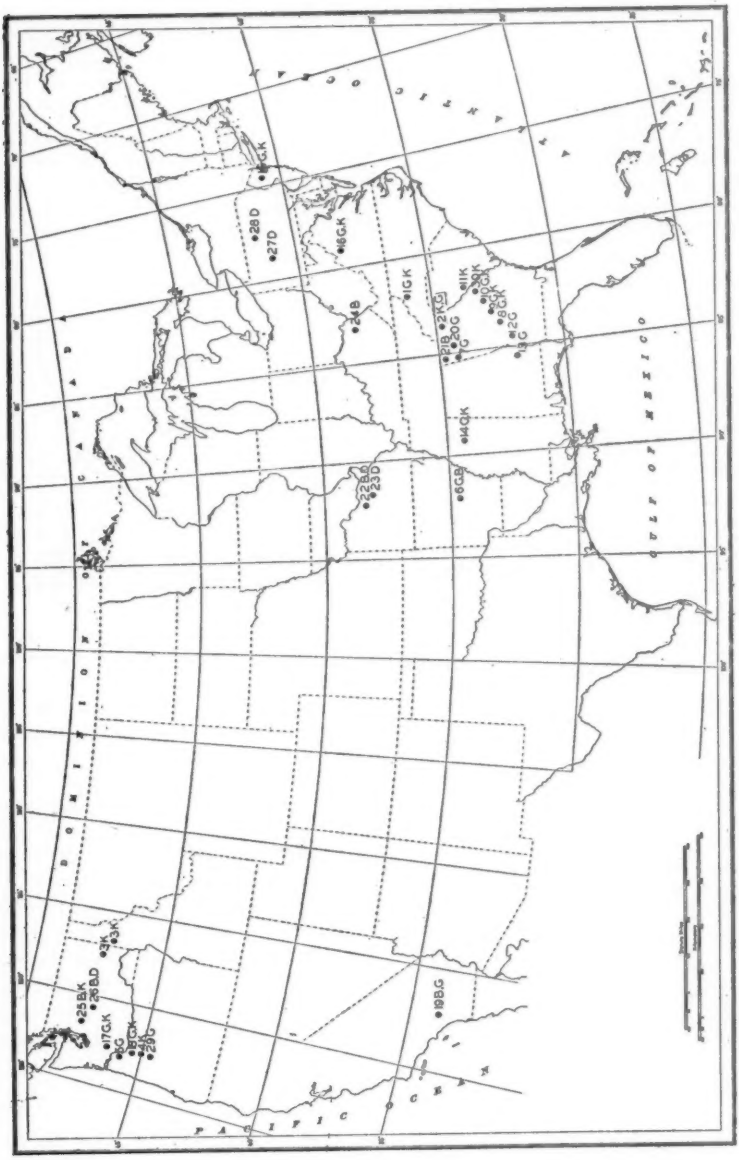
GULF OF MEXICO
PACIFIC OCEAN
DOMINION OF CANADA
1G K
2 8 D
2 7 D
6 G K
2 4 B
1 1 G K
2 1 B
2 0 G
1 1 K
3 0 K
1 0 K
8 K
8 C K
1 2 G
9 C
1 4 G K
6 G B
2 2 B D
2 3 D
3 K
2 5 B K
2 6 B D
1 7 G K
2 G
1 8 G K
2 9 D
1 9 B G

FIGURE 1

siders the oldest terrace level Tertiary. This concept is shared by Parker (1946), who mapped the kaolin deposit as a part of the high-alumina clay program of the U. S. Geological Survey.

Weathering of the original minerals of the granite has been selective. The structural control afforded by increased permeability, such as exists along fractured quartz veins, allowed ground-water solutions to decompose minerals at depths greater than those in less favorable parts of the deposit. Locally, decomposition extends downward 100 feet or more, but the commercial deposits of kaolin are limited to a depth of about 40 or 50 feet, because an appreciable quantity of partly decomposed rock appears at that depth. Orthoclase, which is sparsely represented, decomposes to kaolinite most easily. Microcline, albite, muscovite, and quartz resist decomposition and remain unaltered or only slightly altered even when the surrounding minerals are completely changed to kaolin. Biotite alters to hydromica, and garnet to brownish spots containing iron or manganese oxides.

Several investigators (Somers, 1921; Ries, 1927) assume an isomorphous gradation between kaolinite and muscovite which is marked by a gradual loss of potash and addition of water, whereby in the process of weathering hydromica may represent a transitional stage toward kaolinite. Hickling (1909) studied the China clay of Cornwall, England, where kaolinite and mica make up 90 per cent of the mass and the relative amounts of kaolinite, hydromica, and muscovite are difficult to estimate. Emphasis is placed on the observation of prisms composed of mica at one end and kaolinite at the other, and on the experiment of Johnstone (1889), who found that muscovite changed to hydromuscovite after immersion for 12 months in water saturated with carbon dioxide. Galpin (1912) reports "ribs" of higher index and birefringence intergrown with those of kaolinite and showing practically every gradation between kaolinite and muscovite. Ross and Kerr (1931) found a mineral with moderately high index and birefringence intergrown with kaolinite and considered it was similar to kaolinite except that it contained one molecule of water instead of two. Ample evidence in recent papers demonstrates that the clay minerals initially formed from a specific mineral or rock may differ with

FIGURE 1.—INDEX MAP OF THE UNITED STATES SHOWING LOCALITIES WHERE RESIDUAL AND SEDIMENTARY KAOLIN DEPOSITS (K), GIBBSITE (G), BOEHMITE (B), AND DIASPORE (D) DEPOSITS WERE STUDIED

(1) Spruce Pine district, North Carolina
(2) Rabun County, Georgia
(3) Excelsior, Wash.; Latah County, Idaho
(4) Mehama district, Marion County, Oregon
(5) Ferruginous bauxite in Washington, Columbia and Multnomah counties, Oregon
(6) Bauxite district, Arkansas
(7) Bauxite district, Floyd County, Georgia
(8) Bauxite and kaolin district, Twiggs County, Georgia
(9) Kaolin district, Wilkinson County, Georgia
(10) Kaolin district, Washington County, Georgia
(11) Kaolin district, Aikin district, South Carolina
(12) Andersonville bauxite district, Macon and Sumter counties, Georgia
(13) Bauxite district, Barbour and Henry counties, Alabama
(14) Bauxite district, Pontotoc and Union counties, Mississippi
(15) Kaolin with gibbsite district, Woodbridge, New Jersey
(16) Bauxite district, Augusta County, Virginia
(17) Castle Rock deposit, Cowlitz County, Washington
(18) Molalla deposit, Clackamas County, Oregon
(19) Kaolin deposit, Riverside County, California
(20) Bartow County clay district, Georgia
(21) Walker County clay district, Georgia
(22) Gasconade County district, Missouri
(23) Diaspore district, Phelps, Maries, Crawford counties, Missouri
(24) Olive Hill district, Carter County, Kentucky
(25) Durham district, King County, Washington
(26) Cle Elum deposit, Kittitas County, Washington
(27) Curwensville district, Clearfield County, Pennsylvania
(28) Clinton County clay district, Pennsylvania
(29) Gibbsite deposit, Salem, Oregon
(30) Glascock County clay deposit, Georgia

the conditions and the reacting solutions under which the alteration takes place.

To test Hickling's idea that feldspars weather first to mica and then to kaolinite, numerous specimens of the residual clay and the associated weathered feldspars were examined in oils of known indices of refraction by the immersion method. The most satisfactory materials for this test were selected, and thin sections were made of them and of representative clays. Examination of these thin sections showed orthoclase altered to kaolinite without a suggestion of the formation of intermediate products. Microcline, albite, and parts of perthitic intergrowths are fresh or slightly altered in the same sections where orthoclase and plagioclase containing substantial amounts of the anorthite molecule were completely changed to kaolinite (Pl. 1, fig. 1). The alteration product that locally clouds the surface of the feldspars has low interference colors, which are characteristic of kaolinite rather than of hydromica. Muscovite, which is a primary constituent of the pegmatitic phases of the granite, resists weathering, and unaltered or only slightly altered muscovite is recovered as a by-product during the washing and preparation of the kaolin for utilization. The washed kaolin approaches closely the composition of kaolinite (Table 1) and contains only a small proportion of fine quartz and muscovite.

Two specimens of the clay from Spruce Pine, North Carolina, were submitted to Professor John W. Gruner to determine whether x-ray methods would reveal relict hydromicas too fine to be observed by optical examination. The x-ray patterns indicated that one clay is composed of halloysite and kaolinite, and the other is a mixture of kaolinite, gibbsite, and microcline. The gibbsite recorded in the x-ray pattern is too fine-grained to be visible in thin section. This suggests that finely divided gibbsite is formed by the weathering of the kaolinite, which clouds irregular patches of the microcline perthite. Since finely divided gibbsite should respond first to resilication, it would not be expected to survive when other gibbsite in the specimen had changed to kaolinite or halloysite. The x-ray patterns and petrographic evidence are in agreement that halloysite and kaolinite are the first minerals formed during the initial stages of weathering of the feldspars in these North Carolina deposits. This conclusion agrees with earlier observations of the writer on the feldspars in the granitic rocks of the Sierra Nevada of California (Allen, 1929).

A "hydromica" from Waynesville, North

TABLE 1.—CHEMICAL COMPOSITION OF WASHED KAOLIN, HYDROMICA, AND A GIBBSITE-KAOLINITE-FELDSPAR SPECIMEN

	1	2	3	4	5	6
SiO_2	47.94	46.18	46.5	41.98	40.79	21.65
Al_2O_3	37.02	38.38	39.5	29.54	29.98	49.65
Fe_2O_3	0.60	0.57		9.81	8.07	0.85
FeO	—	—		0.67	2.48	—
TiO_2	0.02	0.04		1.20	1.28	none
CaO	0.30	0.37		none	0.45	none
MgO	0.07	0.42		1.77	2.71	none
Na_2O	0.06	0.10		0.36	0.38	0.21
K_2O	1.25	0.58		2.06	3.47	3.33
Others		0.08				
H_2O − 110°C	13.03	13.28	14.0	1.12	1.20	0.40
H_2O + 110°C				11.32	9.34	23.73
Total	100.29	100.00		99.83	100.15	99.82

1, 2. Washed kaolin, Spruce Pine, North Carolina. Analyses furnished by Harris Clay Company and Kaolin, Inc. (Hunter, 1940).

3. Theoretical composition of kaolinite to satisfy formula $Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$.

4. Hydromica with a little kaolinite. Alteration product of biotite, Lunday mine near Micaville, North Carolina. Analyst, Joseph J. Fahey.

5. Hydromica, Herren property, Waynesville, North Carolina. Analyst, George Steiger (Bayley, 1925).

6. Gibbsite-kaolinite-feldspar specimen derived by weathering of feldspar. W. E. Norton property, Land Lot 20, District 3, Rabun County, Georgia. Analyst, Joseph J. Fahey.

Carolina (Bayley, 1925), appears to have resulted from the alteration of muscovite. The chemical composition of this specimen determined by George Steiger (Table 1, No. 5) is similar to that of a "Hydromica" collected by the writer from the Lunday mine, near Micaville, North Carolina, and analyzed by Joseph J. Fahey (Table 1, No. 4). The hydromica from the Lunday mine was derived by the weathering of biotite, as shown by some of the less altered sheets of biotite, which retain their brownish color near the center and grade outward into bronze to cream-colored hydromica. Sheets uniformly colored bronze were selected by the writer for chemical analysis and for x-ray examination by Professor John W. Gruner, who reported they are composed of hydromica with a little kaolinite. Since the hydromica was embedded in white kaolinite derived from the weathering of feldspars, some kaolinite might have been included between sheets, but none could be detected. It appears that some of this hydromica derived from biotite has changed to kaolinite. Anauxite, which is isomorphous with kaolinite, has been described by the writer as an alteration product of biotite in California (1928).

The optical properties of the hydromica from the Lunday mine are: $\alpha = 1.572$; $\gamma = 1.618$; $\gamma - \alpha = 0.046$; 2V approximately 0°; optical character, negative; some sheets contain spots of yellow iron oxides. The optical properties of the hydromica from Waynesville were not recorded by Bayley, but, because its iron, potassium, and magnesia are higher and its water content is lower, one would expect its indices of refraction to be slightly higher than those determined by the writer for the hydromica from the Lunday mine. Furthermore, since the indices of refraction of the hydromica from the Lunday mine are higher than those of any muscovite listed by Larsen and Berman (1934), it is unlikely that the hydromica from Waynesville could have resulted from the alteration of muscovite, because decrease of potassium and iron and increase of water would decrease rather than increase the indices of refraction of the resulting hydromica. For this reason and because muscovite remains stable and even in tiny sheets retains its characteristic indices of refraction and large optic angle, the writer believes that the hydromica from Waynesville was derived from biotite rather than muscovite.

Rabun County, Georgia

Specimens of microcline-perthite pegmatite from a trench 19 to 22 feet below the surface at the W. E. Norton property, Land Lot 20, District 3, near Clayton, Rabun County, Georgia, were furnished for this investigation by W. H. Yoho (Fig. 1, loc. 2). Most of these consisted of fresh pink microcline-perthite, but on some specimens tiny veinlets and pockets of white powder were surrounded by fresh feldspar. Some of the powder was entirely halloysite or kaolinite; other parts were a mixture of these minerals with gibbsite or feldspar. Two of these powders were submitted to Professor John W. Gruner, who reported that x-ray patterns indicated one was about half halloysite and half gibbsite and the other was crystalline gibbsite with a little microcline and kaolinite. The latter specimen consisted of spherical groups of radiating crystalline gibbsite up to an eighth of an inch in diameter surrounded by fine-grained kaolinite that formed small patches in the microcline-perthite. Several of the spherical masses were isolated and submitted to Joseph J. Fahey for a chemical analysis.

Calculation of this analysis (Table 1, No. 6) indicates 60.5 per cent of gibbsite. If the K_2O and Na_2O in this analysis is combined with enough Al_2O_3 and SiO_2 to form microcline (K_2O, Na_2O, $Al_2O_3 \cdot 6SiO_2$), the sample contains 20.91 per cent microcline. If the remaining silica (8.10 per cent) is combined with sufficient water and alumina to form kaolinite ($Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$), the sample would contain 17.56 per cent kaolinite. The remaining Al_2O_3 (38.96 per cent) and the remaining H_2O (21.54 per cent) would combine to form 60.5 per cent gibbsite ($Al_2O_3 \cdot 3H_2O$). If these last figures are recalculated on the basis of 100 per cent, the results approach closely the 65.4 per cent Al_2O_3 and the 34.6 per cent H_2O required by the formula of gibbsite. X-ray, optical, and chemical data agree that this specimen is a mixture of gibbsite, kaolinite, and microcline. Petrographic observations indicate that kaolinite and halloysite are formed as the first weathering product of the feldspar. In some of these specimens, the change ended with the formation of these kaolin minerals; in others, gibbsite was formed from the kaolin minerals by desilication. The radiating groups of crystalline

gibbsite have sharp contacts with the surrounding fine-grained kaolinite and were developed by growing outward at the expense of the kaolinite. This process probably took place during an old incomplete erosion cycle on the rocks of the Piedmont plateau rather than during the recent cycle, but the exact age of the weathering is unknown. Most of the weathered feldspathic rocks of the Piedmont plateau examined by the writer contain only clay minerals, which dot the feldspars and extend as a network across them.

Columbia River Region of Oregon, Washington, and Idaho

In the Columbia River region of Oregon, Washington, and Idaho (Fig. 1, loc. 3), plagioclase of basaltic lavas is altered to kaolinite (Pl. 1, fig. 2), halloysite, endellite (Pl. 1, fig. 3), and nontronite. However, nontronite is also formed from augite, iddingsite, basaltic glass, and palagonite under conditions of poor drainage and in the presence of magnesium. The kaolin minerals, kaolinite and halloysite, are formed under good drainage and neutral or slightly acid conditions. Halloysite and kaolinite occur in the same vertical section, which suggests they were formed in this region under nearly similar conditions of weathering. In some specimens, plagioclase is altered along cleavages and cracks to clay minerals that extend as irregular zones across the crystals (Pl. 1, figs. 2, 3). All stages from partly altered to completely kaolinized plagioclase have been

PLATE 1.—PHOTOMICROGRAPHS OF ALTERED FELDSPARS

FIGURE 1. ORTHOCLASE AND PLAGIOCLASE COMPLETELY ALTERED TO KAOLINITE (K, DARK). MICROCLINE (M, WHITE) UNALTERED

Kaolin Inc. mine, 5 miles S. E., Spruce Pine, N. C.

FIGURE 2. EARLY STAGE OF ALTERATION OF PLAGIOCLASE TO KAOLINITE (K)

(K) cuts plagioclase laths as veinlets and embayed areas leaving relict areas of plagioclase (white) in a weathered basalt. Nontronite (N) formed from basaltic glass, Ilmenite (I). Drill hole 19, depth 58 feet, Excelsior clay deposit, Washington.

FIGURE 3. EARLY STAGE OF ALTERATION OF PLAGIOCLASE (P) TO ENDELLITE ALONG CLEAVAGES AND VEINS

Augite, iddingsite, and ilmenite are unaltered. Weathered basalt, 8 miles south of Salem, Oregon.

FIGURE 4. LATHS OF FELDSPAR (F) PARTLY ALTERED TO AND CUT BY VEINLETS OF HALLOYSITE

Areas of halloysite (H) fill space between feldspars. Altered syenite. Bauxite district, Arkansas.

FIGURE 5. EARLY STAGE OF WEATHERING OF AN AEGIRITE SYENITE

Feldspar (F) is partly altered to halloysite white, but aegirite retains original optical properties. Bauxite district, Arkansas.

FIGURE 6. LATH OF FELDSPAR ALTERED TO KAOLINITE (K) AND GIBBSITE (DARK G) PARALLEL TO THE CLEAVAGE

Center of boulder surrounded by bauxite. Neilson underground mine, Bauxite district, Arkansas.

PLATE 2.—PHOTOMICROGRAPHS OF SEDIMENTARY CLAYS

FIGURE 1. PISOLITE IN WHICH KAOLINITE PLATES SHOW "WORM" STRUCTURE

Soft kaolin. Paragon Mines, Langley, South Carolina.

FIGURE 2. "PISOLITE" OF CLAY REWORKED FROM AN UNDERLYING FORMATION

Quartz grains are more abundant near the margin that those in the matrix or near the center of the "pisolite". At top small veins of gibbsite (G) cut the edges of the "pisolite". Drill core of hole 29, Pinedale, Mississippi.

FIGURE 3. PISOLITE FORMED BY CONTRACTION CRACKS AS CLAY DRIES

Open spaces are developed (O, white) and gibbsite forms in and near them. Andersonville bauxite district, Georgia.

FIGURE 4. PISOLITES FORMED BY CONTRACTION IN BAUXITE WITH 67 PER CENT Al_2O_3

Gibbsite is concentrated along the cracks. Depth 52 feet. Andersonville bauxite district, Georgia.

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Castle Rock, Washington.

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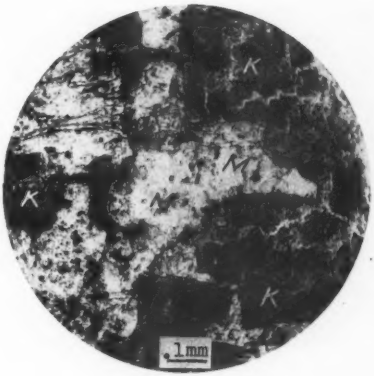


FIGURE 1

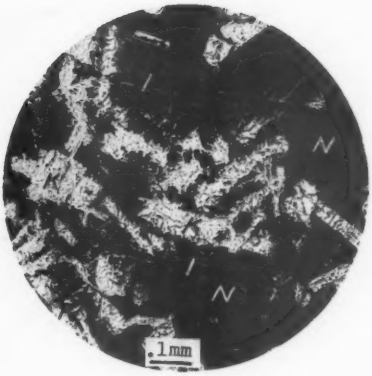


FIGURE 2

FIGURE 3

FIGURE 4

FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS OF ALTERED FELDSPARS

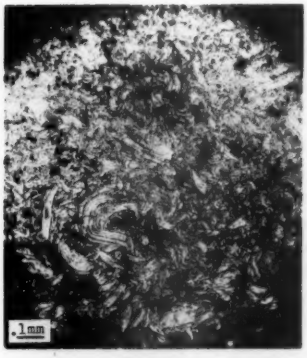


FIGURE 1

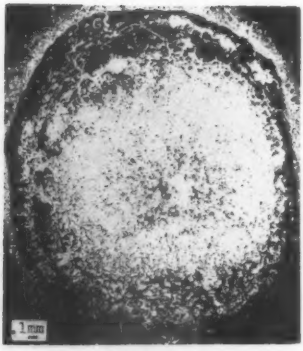


FIGURE 2

FIGURE 3

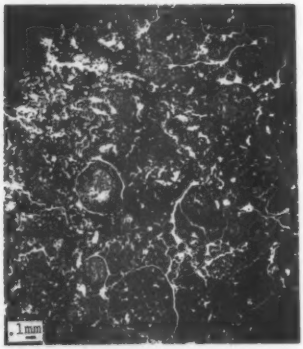


FIGURE 4

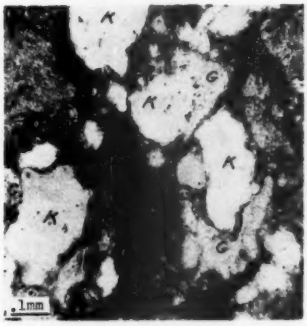


FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS OF SEDIMENTARY CLAYS

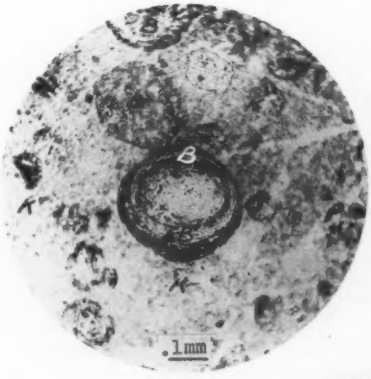
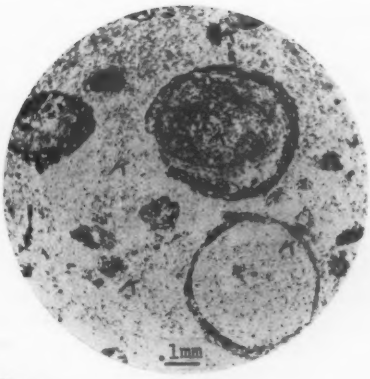


FIGURE 1



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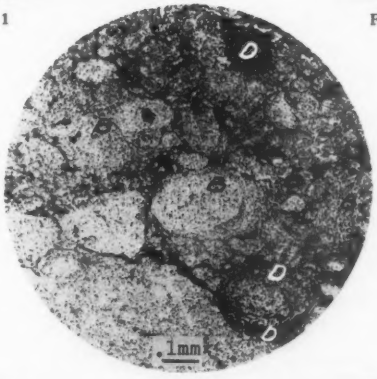
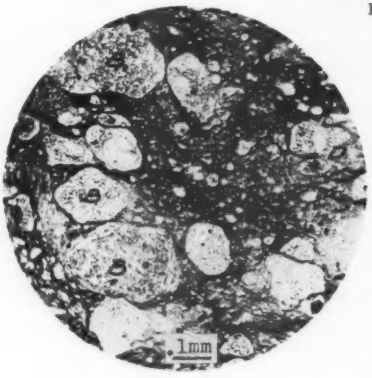


FIGURE 3



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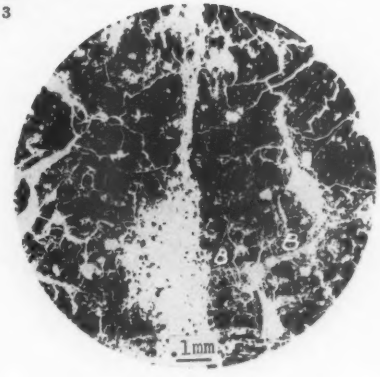

FIGURE 5

PHOTOMICROGRAPHS OF BOEHMITE

FIGURE 1

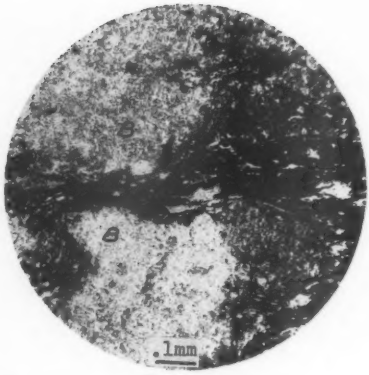

FIGURE 2

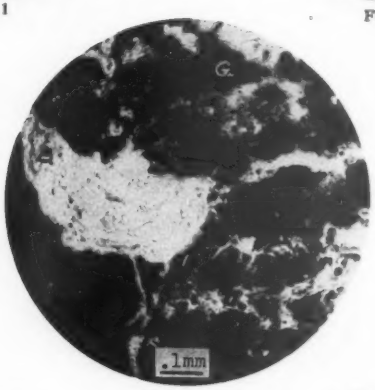

FIGURE 3

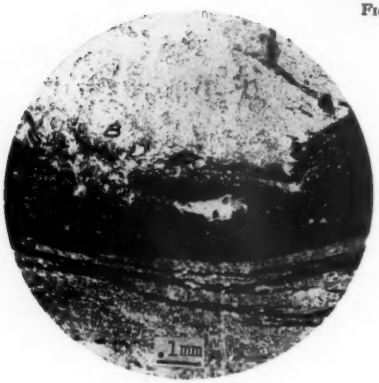

FIGURE 4

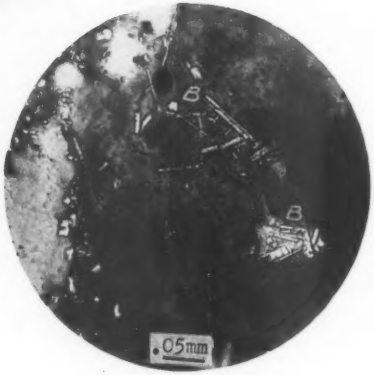

FIGURE 5

PHOTOMICROGRAPHS OF BOEHMITE

observed. Many specimens were examined in thin sections or in oils by the immersion method, but no traces of gibbsite or hydromica were found in them. In the Mehama district, Marion County, Oregon (Fig. 1, loc. 4), the original feldspar phenocrysts of porphyritic volcanic rocks have been altered to kaolinite with the preservation of relict twinning. At the outcrops, pseudomorphs of kaolinite retain all the crystal faces of the original feldspar phenocrysts. The relations in the Columbia River region indicate that clay minerals were the first products formed during the initial stages of weathering of plagioclase, ferromagnesian minerals, and basaltic glass (Allen and Scheid, 1946).

In the lower part of the ferruginous bauxite (also known as laterite) deposits in Washington, Columbia, and Multnomah counties, Oregon, similar clay minerals occur in the same relations (Fig. 1, loc. 5). It is reasonable to believe that they were formed under the same conditions of weathering. Locally these clay minerals form a transitional layer over 100 feet thick separating the bauxite at the top of the deposit from the underlying parent basaltic rocks of Miocene age. The shapes of the minerals, the vesicles, and other structures of the basaltic lavas are retained, but their compositions are changed to those of clay minerals requiring a loss of silica, calcium, mangesium, iron, and alkalies. A continuous series of core samples with chemical analyses at each 2-foot interval contributed by Alcoa Mining Company (Allen, 1947; 1948) shows that weathering penetrated downward more than 175 feet; as weathering proceeded, the content of alumina, ferric oxide, titania, and water was increased progressively but not regularly as silica was removed by desilication (Fig. 2). At a depth of 67 feet, plagioclase laths are altered to gibbsite at the edges, but some halloysite is present near the center. This suggests that halloysite was the first weathering product of the plagioclase, as has been observed in scores of specimens from the Columbia River region, and that halloysite changed to gibbsite by loss of silica. Also, at depths from 65 to 20 feet at this deposit, nontronite derived from basaltic glass and ferromagnesian minerals lost silica and changed to gibbsite along veins and embayed areas in the nontronite. The utilization of the alumina in the nontronite and its fixation as gibbsite liberated the iron from nontronite as oxides. Some of the iron oxides was deposited with the gibbsite, staining it or forming the pisolites of

PLATE 3.—PHOTOMICROGRAPHS OF BOEHMITE

FIGURE 1. BOEHMITE (B) REPLACING OÖLITES OF KAOLINITE IN A FLINT CLAY
Burnt House mine, Olive Hill district, Carter County, Kentucky.

FIGURE 2. OÖLITES COMPOSED ENTIRELY OF KAOLINITE (K) IN FLINT CLAY
Olive Hill district, Carter County, Kentucky.

FIGURE 3. BOEHMITE (B) REPLACING OÖLITES OF KAOLINITE IN FLINT CLAY. DIASPORE (D) REPLACING SOME OÖLITES
Swiss, Gasconade County, Missouri.

FIGURE 4. BOEHMITE (B) ASSOCIATED WITH DIASPORE (D) REPLACES NODULES OR PELLETS AT TOP OF IRON DEPOSIT
Cle Elum, Kittitas County, Washington.

FIGURE 5. BOEHMITE ASSOCIATED WITH GIBBSITE IN A PISOLITE
Alberhill, Riverside County, California.

PLATE 4.—PHOTOMICROGRAPHS OF BOEHMITE

FIGURE 1. BOEHMITE (DARK, B) REPLACING KAOLINITE ALONG CLEAVAGE IN A KAOLINITE "BOOK" IN SHALES OF THE PUGET GROUP
Durham, King County, Washington.

FIGURE 2. BOEHMITE (B) REPLACING CARBONACEOUS STREAK IN SHALES OF THE PUGET GROUP
Durham, King County, Washington.

FIGURE 3. BOEHMITE (B) ASSOCIATED WITH GIBBSITE (G) IN A BLACK PISOLITE
Near Berger, Pulaski County, Arkansas.

FIGURE 4. BOEHMITE (B) REPLACING CARBONACEOUS LAYER NEAR COAL IN SHALE OF THE PUGET GROUP
Kummer, King County, Washington.

FIGURE 5. BOEHMITE (B) CRYSTALS IN SHALE OF THE PUGET GROUP
Kummer, King County, Washington.

goethite, hematite, or maghemite that characterize this high-iron bauxite. All the features of the bauxite deposits of Oregon can be explained by a continuation of the same process of weathering and thorough leaching that formed the kaolin minerals. This weathering has been described as a two-stage process (Allen, 1948) to emphasize (1) that clay minerals are the first products formed, and (2) that only locally in this region has desilication reached a stage where gibbsite and high-iron bauxite are formed from the clay minerals.

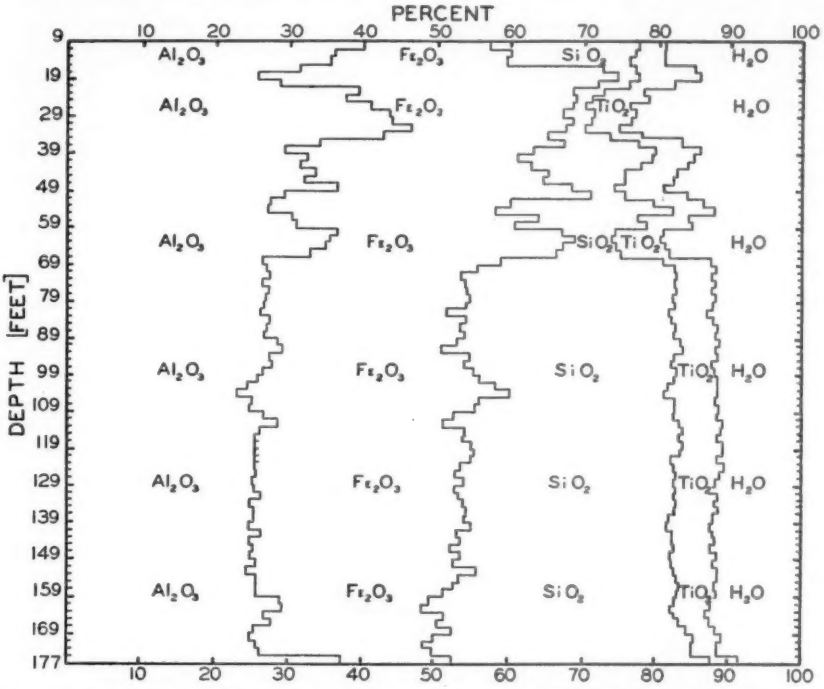

FIGURE 2.—VARIATION IN CHEMICAL COMPOSITION WITH DEPTH IN A BAUXITE DEPOSIT IN OREGON DERIVED FROM BASALTIC ROCKS

Analyses at 2-foot intervals by Alcoa Mining Company.

Bauxite District, Arkansas

Five specimens of altered syenite collected by the writer in the Arkansas bauxite district (Fig. 1, loc. 6) show halloysite or kaolinite invading feldspar laths; the relations suggest that the kaolin minerals, halloysite or kaolinite, are the initial product of weathering of the feldspars. Specimens 1, 2, and 3 are from an open-pit mine of the Republic Mining and Manufacturing Company (now Alcoa Mining Company) near Bauxite, Arkansas. In specimen 1, some parts of the feldspar are fresh enough to retain their characteristic optical properties, but others are changed along cleavages, at the center, or the edges to an isotropic mineral with the optical properties of halloysite. In places the matrix around the feldspars is converted to halloysite stained slightly yellow with iron oxides. A complex series of irregular wavy veinlets of halloysite cuts across the feldspars and the remains of other minerals except the ilmenite in this thin section (Pl. 1, fig. 4). Professor John W. Gruner reported that the diffraction pattern of this specimen indicates halloysite but no gibbsite.

In specimen 2, from another part of the same open pit, cleavage sheets of kaolinite form

characteristic "books" or long winding "worms" of varying width that have the optical properties of kaolinite.

Specimen 3 is from the center of an angular block of syenite about 3 or 4 feet in each dimension, is enclosed in bauxite, and is exposed along the walls of an open pit. In thin section, halloysite is seen to have formed from feldspar but has left parts of the feldspar unaltered. This alteration of the felsdpar laths and the groundmass to halloysite represents an early stage of weathering, because the aegirite (an easily altered mineral) in the thin sections retains its original green color (Pl. 1, fig. 5), pleochroism, and birefringence.

In specimen 4, from the center of another angular block in the open pit mine of the American Cyanamid and Chemical Company near Berger (NE1/4 sec. 9, T, 1 S., R. 12 W.), alteration has gone further, and all the minerals in the thin section except ilmenite are changed to halloysite. The shapes of the feldspars and traces of the outline of the ferromagnesian minerals are preserved. Both these blocks are surrounded by bauxite of commercial grade composed of gibbsite. The contact of the blocks, which retain the granitoid texture of the original syenite, with the surrounding bauxite is sharp, and the blocks contrast with the surrounding bauxite.

No. 5, from the Neilson underground mine (sec. 23, T. 2 S., R. 14 W.), consists of a series of six specimens from the center of a boulder through a zone 1½ feet thick to the surrounding commercial-grade bauxite. The thin section of the innermost sample is mostly kaolinite with a little gibbsite (Pl. 1, fig. 6). This is confirmed by an x-ray pattern of the material examined by Joseph M. Axelrod. Either fine-grained kaolinite or halloysite is present in all six thin sections. In thin sections of the outermost specimens, crystalline gibbiste predominates, and a definite but not regular increase in the amount of gibbsite can be seen toward the outer edge. In the center of the boulder, the feldspars are now mostly kaolinite, but in the outer zone kaolinite is irregularly distributed in thin bands parallel to the cleavages, and gibbsite increases in amount at the expense of kaolinite. Where a vein of halloysite cuts through the edge of the boulder, concentric cracks in the halloysite are filled with gibbsite that extends as tonguelike projections into the halloysite. Nearly complete rings of gibbsite that enclose halloysite are formed in places. At the outermost edge on the lower side of the boulder, a nodule of siderite has been formed. In it are coarse plates of kaolinite, and in some of them siderite has entered and formed wedges that either force the sheets apart or replace them.

The long lath-shaped feldspars in specimens 1, 3, and 5 are like those forming the texture of the so-called "granitic" bauxite. The shapes of the feldspars could be preserved best during the process of desilication, when the removal of silica and alkalies allowed adjustment of volume without distortion. The presence of kaolinite or halloysite in these specimens of weathered syenite of various textures suggests that desilication of kaolinized syenite to form gibbsite and bauxite has taken place in Arkansas. This process has been questioned by Goldman and Tracey (1946, p. 575). A detailed study of the region should be undertaken to determine quantitatively the importance of the kaolinization of nepheline syenite as an intermediate step in its weathering to bauxite as contrasted with the formation of gibbsite directly from the syenite.

Desilication of Sedimentary Clays

Evidence from Texture and Structure of Clays

The preservation of textures and structures of sedimentary clays in the aluminum hydrate materials suggests that gibbsite, boehmite, or diaspore were formed from the clay minerals after their deposition. Many residual and colluvial clays lack similar characteristic primary structures that could serve as points of reference in tracing the effects of desilication that may attack them as well as sedimentary clays. The primary structures of sedimentary clays are developed during deposition and include: stratification, colloidal forms, oölites, pellets, and "spotted clays". These "spotted clays" result from small, local accumulations or spots of carbonaceous clay.

The stratification of clays is not conspicuous because the constituent particles are very fine and occur in aggregates of random arrangement. However, in some clays stratification is indicated by the alignment of plates or sheets

of the clay minerals and is emphasized where carbonaceous materials or color bands are present. The term "shale" is reserved for those clay rocks that can be divided easily into thin laminae, which are connected with the lamination imparted by the micaceous minerals arranged nearly parallel to the bedding.

Colloidal materials in clays assume various shapes that can be seen best in thin sections. Some are irregular and resemble smears, but others are concentric. Fanlike forms consist of curved bands that widen away from the converging end.

Oölites are present in many clays and consist of spherical or ellipsoidal particles with an internal concentric structure. Particles of the same shape and internal structure but larger than 2 mm in diameter are called pisolites. Oölitic structure in clays originates in several ways. Some that have been suggested are:

1. Direct precipitation of suspensoids in the basin of deposition as aggregates with concentric structure.

2. Rearrangment and adjustment of colloidal particles around a point during or shortly after deposition.

3. Differential shrinkage of the area of the future oölite with respect to the surrounding clay, from which it differs in mineral composition, particle size, stacking of the molecular sheets, and plasticity. The writer has seen concentric cracks, outlining areas with the dimensions of oölites and pisolites, form in clays during drying. The oölitic area so formed is composed of clay that is softer and more plastic than that surrounding the open concentric crack (Pl. 2, fig. 3).

4. Incorporation of fragments derived from an older clay formation and rolled about until rounded. So-called pisolites in the Wilcox formation of Tennessee and Mississippi contain more quartz grains than the surrounding clay. Some of these were probably derived from the underlying Porters Creek clay. Others contain veins of gibbsite cutting across the rims of the "pisolites", indicating that the "pisolites" were fully developed when gibbsite formed. (Pl. 2, fig. 2).

The term "pellet" has been applied to ovate masses of clay 0.05–15.0 mm long arranged with their long axes parallel to the bedding (Allen and Nichols, 1945). At Hobart Butte, Oregon, the pellets are surrounded by clay either containing organic matter or stained red, yellow, or brown with iron oxides. Diatoms and fragments of charcoal are present in the matrix. The pellets are molded against quartz and rock fragments, indicating that they were soft at the time of deposition. The pellets are of interformational origin, composed of clay fragments broken from thin clay layers that had dried on the flood plains of streams and were rounded by transportation.

In "spotted clays", the spots vary in size, shape, and color. Some are small and irregular; others are ovate like oölites or pellets. Carbonaceous materials are present in all of them; this suggests that their formation is connected with the associated carbonaceous material. Boswell (1930) reports experiments in which the coprecipitation of colloids in fine sediments was facilitated by the presence of humic acids, and the rapidity of fall was dependent on the concentration. During flocculation and settling, grains were entrapped and formed a mixture of carbonaceous material and clay.

Replacement by Gibbsite

Georgia and Alabama.—In Georgia and the neighboring States of the Coastal Plain, oölites and pisolites composed of kaolin minerals occur in many clays that contain no aluminum hydrate minerals and show no sign of alteration. The oölites are so widespread and resemble the surrounding matrix so closely in mineral composition that they appear to have been formed by sedimentation. The writer has observed them in the clays associated with the bauxite-filling depressions in the Paleozoic rocks of Floyd and Bartow counties, Georgia (Watson, 1904) (Fig. 1, locs. 7, 20), in the Tuscaloosa formation of Cretaceous age in Twiggs (Fig. 1, loc. 8), Wilkinson (Fig. 1, loc. 9) and Washington counties (Fig. 1, loc. 10), Georgia (Smith, 1929), and Aiken County, South Carolina (Lang, 1940) (Fig. 1, loc. 11), and the Wilcox group of Eocene age in Macon and Sumter counties, Georgia (Shearer, 1917) (Fig. 1, loc. 12). The bauxite and kaolin deposits of Wilkinson County, Georgia, were mapped by Warren and Thompson (1943), the kaolin deposits of Washington and Twiggs counties, Georgia, by

Thompson (1943a; 1943b), the bauxite and clay deposits in the Andersonville district, Georgia, by Zapp (1943), and the bauxite deposits of Barbour and Henry counties, Alabama (Fig. 1, loc. 13), by Warren (1942).

A detailed correlation of the formations of Tertiary age in the eastern Coastal Plain of the United States has been prepared by MacNeil (1946; 1947). A generalized sequence of formations in the bauxite areas of the coastal plain of Georgia and Alabama is given here for convenience.

Alabama		Georgia
Western	Eastern	
Tertiary	Tertiary	Tertiary
Eocene	Eocene	Eocene
Jackson group	Jackson group	Jackson group
Claiborne group	Claiborne group	Claiborne group
Wilcox group	Wilcox group	Wilcox group
Paleocene	Paleocene	Paleocene
Midway group	Midway group	Midway group
Naheola formation		
Porters Creek clay		
Clayton formation	Clayton formation	Clayton formation
Cretaceous	Cretaceous	Cretaceous
Upper Cretaceous	Upper Cretaceous	Upper Cretaceous
Selma group	Selma group	
Prairie Bluff chalk	Providence sand	Providence sand
Ripley formation	Ripley formation	Ripley formation
Demopolis chalk		Cusseta sand
Mooreville chalk	Blufftown formation	Blufftown formation
Eutaw formation	Eutaw formation	Eutaw formation
McShan formation	?	
Tuscaloosa group	Tuscaloosa formation	Tuscaloosa formation

COMPOSITION AND STRUCTURE OF KAOLINS: The writer has studied petrographically the clays and associated bauxite in each of these districts. The chief mineral in all the clays which can be correctly termed "sedimentary kaolins" is kaolinite, but some halloysite may be present in the finely divided isotropic parts. Small amounts of quartz, muscovite, zircon, rutile, tourmaline, and magnetite are locally present. Petrographic, x-ray, and differential-thermal analyses indicate that the varieties known as "hard," "flint," "semihard," and "soft" do not differ essentially in mineral composition, but the kaolinite in the "soft" clays is present as large, well-oriented sheets arranged in "books" or "worms" that are lacking in the other types. The kaolinite of the "hard" and "flint" types contains colliform structures in which concentric, layered areas of low birefringence form compact aggregates. These relationships were first observed in thin sections of clays at Dry Branch and Andersonville, Georgia, studied in September 1941. Samples of "soft" kaolins from Sandersville and McIntyre and "hard" kaolins from Oconee and McIntyre, Georgia, were also obtained from the U. S. Bureau of Mines Experiment Station at Tuscaloosa, Alabama, through the courtesy of T. A. Klinefelter, who supplied notes on the clay types and the localities by R. W. Smith. The same relations as well as the tendency of quartz grains to occur abundantly in the "hard" kaolins were observed in these specimens. The writer concluded that "soft" clays are composed of coarser kaolinite sheets than the "hard" types. Independently the same conclusion was reached during two other investigations of the kaolins of Georgia (Mitchell and Henry, 1943; Klinefelter *et al.*, 1943).

During this petrographic study the writer was impressed with the importance of the variation in grain size, colloform structures, compaction, and the physical properties of clay masses in controlling the porosity and permeability of the clays both horizontally and

vertically and thus providing paths for groundwater solutions.

The clay associated with bauxite in Floyd County, Georgia, has the chemical composition shown as No. 3 (table 2); the clay below the bauxite near Gordon, Wilksinson County, Georgia, is shown as No. 6; and the clay below the bauxite at the Sweetwater Bauxite mine, Sumter County, Georgia, as No. 12. In these clays the Al_2O_3 content ranges between 37.59 and 41.33 per cent. The only clay mineral indicated by optical and x-ray methods in the clays and oölites is kaolinite.

TABLE 2.—CHEMICAL COMPOSITION OF PISOLITES, MATRIX, AND ASSOCIATED CLAYS OF GEORGIA
Lawrie H. Turner, analyst

	1	2	3	4	5	6	7	8	9	10	11	12
SiO_2	6.13	2.90	41.08	3.93	34.73	39.46	4.14	22.95	21.38	37.69	29.94	44.78
Al_2O_3	60.74	60.87	41.33	65.10	44.52	41.03	61.77	51.39	52.19	41.72	47.10	37.59
Fe_2O_3	0.73	3.14	0.63	0.70	1.15	0.72	1.05	1.55	1.35	1.00	0.90	1.04
TiO_2	0.25	1.05	0.80	1.30	0.50	1.35	2.10	1.45	1.50	1.35	1.50	1.65
CaO	0.25	0.12	0.80	0.20	0.06	0.21	0.07	0.04	0.00	0.12	0.15	0.19
MgO	0.41	0.32	0.10	0.30	0.00	0.32	0.00	0.40	0.50	0.35	0.40	0.43
Moisture	0.95	0.44	0.37	0.97	0.97	0.86	0.68	0.67	0.51	0.62	0.76	0.64
Ignition loss	30.54	31.16	15.61	27.50	18.07	16.05	30.19	21.55	22.57	17.15	19.25	13.68

1. Matrix around hard pisolites (No. 2).
2. Hard pisolites in a hard matrix (No. 1).
3. Clay from same pit as Nos. 1 and 2.

1, 2, 3, New Prospect Baptist Church mine, 14 miles southwest of Rome, Land Lot 837, District 3, Section 4, Floyd County, Georgia.

4. Pisolites from matrix (No. 5).
5. Matrix around pisolites (No. 4).
6. Clay below pisolites (No. 4) with only a slight suggestion of pisolitic structure.

4, 5, 6, Fitzpatrick Place, 3 miles south of Gordon, Wilkinson County, Georgia.

7. Pisolites from bauxite.
8. Firm pisolites from upper part of bauxite.
9. Matrix around pisolites (No. 8).
10. Soft pisolites, 1 ft. below No. 8.
11. Matrix around soft pisolites No. 10.
12. Clay with no pisolites, 2 ft. below No. 9, 7–12. Sweetwater Bauxite mine, Sumter County, Georgia.

Analyses courtesy Georgia Division of Mines and Geology.

COMPOSITION AND STRUCTURE OF BAUXITE: Gibbsite is present in clays or in bauxite of the Coastal Plain of Georgia in which the alumina content is higher than 41 per cent. Where the gibbsite grains are sufficiently large for petrographic observation, the textures suggest that the gibbsite has formed from the clay minerals. In some nonpisolitic bauxite (No. 1, Table 3), gibbsite is so fine-grained that it shows no interference colors, appears opaque by reflected light, and resembles leucoxene optically. Gibbsite particles of submicroscopic size are present in many kaolin deposits. Small amounts of gibbsite produce an opaque effect and escape detection by ordinary petrographic observation.

Some bauxite with pisolites has a higher silica content and lower alumina than the bauxite with no visible pisolites. (*Compare* No. 4, Table 3, with No. 1, Table 3.) The quality of a bauxite sample cannot be accurately estimated from its appearance. Optical determinations profit by checking by x-ray and differential-thermal methods, and a chemical analysis furnishes a useful control. A portable differential-thermal analysis unit for bauxite exploration has been designed and described by Hendricks *et al.* (1946).

At three bauxite localities in Georgia and three in Alabama, the writer collected bauxite samples and later separated the pisolites from the surrounding matrix. The pisolites were analyzed separately from the matrix of each bauxite sample and the results listed in Tables 2 and 3. In general the percentages of alumina, iron oxide, and titania are higher and of the

silica lower in the pisolites than in the corresponding matrix.

In May 1942 large continuous channel samples of the overlying clay, bauxite, and underlying clay were cut from the vertical face exposed in the Sweetwater mine in the Andersonville district, Sumter County, Georgia. These were wrapped so that the interiors retained most of the moisture present at the mine. Observations on this series of specimens in the laboratory extended over several days. During that interval concentric cracks about the size of pisolites formed in some of the samples. The clay in the interiors surrounded by the contraction cracks had the optical properties of kaolinite and so did the clay on the outside of the crack, but the clay in the interior differed from that outside the crack in plasticity and softness. This indicates that the physical properties of small local areas of clay vary sufficiently from the clay around them that shrinkage cracks form and outline a type of pisolite or clay pellet (Pl. 2, fig. 3). Where these are composed of the same minerals, a difference in size of grain or in stacking of the molecular sheets could cause the variation in physical properties, but at other deposits small amounts of halloysite or montmorillonite (Ca, Mg) $O \cdot Al_2O_3 \cdot 4SiO_2 \cdot NH_2O$ mixed with the kaolinite could increase this local variation in physical properties. Cracks of this type were observed in thin sections of the clays associated with other bauxite deposits. At Stevens Pottery, Baldwin County, and at three localities in Wilkinson County, Georgia, some montmorillonite was mixed with the kaolinite in the interior of the contraction area. Also, tiny shreds of montmorillonite were observed in the thin sections of bauxite from Wilkinson and Sumter counties, Georgia.

TABLE 3.—CHEMICAL COMPOSITION OF PISOLITES, MATRIX AND BAUXITE OF ALABAMA
H. S. Hodges, analyst

	1	2	3	4	5	6	7	8
SiO_2	19.57	12.59	11.62	30.86	6.60	6.18	22.80	10.62
Al_2O_3	53.78	56.93	57.94	47.26	59.11	60.74	52.01	58.48
Fe_2O_3	0.50	0.90	1.00	1.24	0.90	0.90	0.60	1.20
TiO_2	0.87	1.48	1.42	0.82	1.27	1.42	1.06	1.10
Ignition loss	24.77	27.87	27.64	19.39	29.80	30.24	22.90	28.35

1. Siliceous bauxite with no pisolites, Baker mine, Barbour County, 13 miles from Eufaula, Alabama.
2. Matrix around pisolites (No. 3). Smith No. 3 mine, Eufaula district, Barbour County, Alabama.
3. Pisolites in matrix (No. 2).
4. Pisolitic clay, Smith No. 3 mine, Eufaula district, Barbour County, Alabama.
5. Matrix around pisolites (No. 6) Alabama Bauxite Company mine, Henry County, 11 miles, Abbeville, Alabama.
6. Soft pisolites in matrix (No. 5).
7. Matrix around soft pisolites (No. 8), Alabama Bauxite Company mine, Henry County, 11 miles, Abbeville, Alabama.
8. Soft pisolites in Matrix (No. 7).

Analyses courtesy Geological Survey of Alabama.

In connection with the co-operative project of the U. S. Bureau of Mines and the U. S. Geological Survey in the Andersonville bauxite district, Georgia, a continuous length of core was furnished the writer for petrographic study. The chemical composition of this core at about half-foot intervals from the kaolin at the top through the siliceous bauxite to the kaolin at the bottom is given as Table 4. These analyses indicate a variation in alumina content of the kaolin at the top from 38.2 to 39.1 per cent, a progressive but irregular increase of alumina with depth to siliceous bauxite with 53.6 per cent alumina, a decrease to 48.0 per cent alumina, and an increase to 54.3 per cent alumina, then the underlying kaolin with 39.9 per cent alumina is reached (Fig. 3). This core was studied petrographically using the immersion method; and 12 thin sections of critical parts were made, studied, and compared with others from the Andersonville bauxite deposit.

Pisolites of gibbsite surrounded by contrac-

TABLE 4.—CHEMICAL ANALYSES OF CORES FROM DRILL HOLE ANG. 18, ANDERSONVILLE DISTRICT, GEORGIA

Depth in feet	Al_2O_3	SiO_2	Fe_2O_3	TiO_2	Ignition Loss*	Total
19.5′–21.0′	39.0	44.4	1.1	1.6	14.2	100.3
21.0′–21.5′	39.1	43.9	0.8	1.4	14.9	100.1
21.5′–22.0′	38.8	44.3	0.7	1.4	14.7	99.9
22.0′–22.5′	38.5	44.4	0.8	1.5	14.6	99.8
22.5′–23.0′	38.3	45.3	0.7	1.4	14.3	100.0
23.0′–23.5′	38.6	44.7	0.6	1.4	14.4	99.7
23.5′–24.0′	38.4	45.0	0.7	1.4	14.6	100.1
24.0′–24.3′	38.9	44.5	0.6	1.4	14.4	99.8
24.3′–25.0′	38.4	44.4	0.6	1.4	14.3	99.1
25.0′–25.5′	38.3	44.8	0.6	1.5	14.1	99.3
25.5′–26.0′	38.8	44.9	0.6	1.5	14.4	100.2
26.0′–26.5′	38.2	44.1	0.6	1.5	14.5	99.9
26.5′–27.0′	40.5	41.2	0.5	1.6	15.2	99.0
27.0′–27.5′	43.0	38.8	0.6	1.6	15.7	99.7
27.5′–28.0′	43.6	37.4	0.6	1.6	16.7	99.9
28 0′–28.5′	44.6	35.8	0.6	1.6	16.0	98.6
28.5′–29.0′	44.7	34.0	0.6	1.7	16.1	97.1
29.0′–29.5′	46.0	32.2	0.6	1.6	17.2	97.6
29.5′–30.0′	47.0	30.2	0.5	1.6	18.1	97.4
30.0′–30.5′	47.3	30.0	0.6	1.6	17.9	97.4
30.5′–31.0′	45.1	30.9	0.5	1.6	18.4	96.5
31.0′–31.5′	46.9	29.0	0.7	1.6	18.8	97.0
31.5′–32.0′	45.8	27.0	0.6	1.8	18.7	96.6
32.0′–32.5′	48.4	27.7	0.8	1.7	18.4	97.0
32.5′–33.0′	47.0	28.3	0.8	1.5	19.0	96.6
33.0′–33.5′	44.8	29.9	0.6	1.6	19.0	95.9
33.5′–34.0′	45.6	30.2	0.8	1.6	18.2	96.4
34.0′–34.5′	47.7	29.4	0.6	1.7	17.5	96.9
34.5′–35.0′	45.6	32.6	0.7	1.6	16.3	96.8
35.0′–35.5′	45.0	33.5	0.7	1.6	16.0	96.8
35.5′–36.0′	43.5	35.6	0.9	1.6	15.7	97.3
36.0′–36.5′	43.0	37.0	0.9	1.6	14.9	97.4
36.5′–37.0′	41.4	39.8	0.8	1.5	14.1	97.6
37.0′–37.5′	41.5	40.1	0.7	1.6	14.2	98.1
37.5′–38.0′	44.5	35.6	0.7	1.6	15.5	97.9
38.0′–38.5′	44.6	35.8	0.7	1.8	15.0	97.9
38.5′–39.0′	43.8	37.0	0.7	1.8	14.6	97.9
39.0′–39.5′	43.4	37.6	0.7	1.8	14.3	97.8
39.5′–40.0′	44.1	37.2	0.7	1.8	14.5	98.3
40.0′–40.5′	43.3	37.0	0.8	1.7	16.2	99.0
40.5′–41.0′	44.1	34.4	0.7	1.8	16.6	97.6
41.0′–41.5′	47.6	27.8	0.8	2.1	19.1	97.4
41.5′–41.75′	45.9	30.9	0.8	1.9	18.7	98.2
41.75′–42.0′	50.6	23.2	0.7	2.1	21.1	97.7
42.0′–42.5′	50.4	22.8	0.8	2.1	21.0	97.1
42.5′–43.0′	52.6	19.8	0.8	1.9	22.4	97.5
43.0′–43.5′	53.6	17.7	0.8	2.4	22.9	97.4
43.5′–44.0′	51.8	21.6	0.7	2.2	21.6	97.9
44.0′–44.5′	48.0	27.8	0.7	2.1	19.1	97.7
44.5′–45.0′	49.9	23.3	0.7	2.4	21.4	97.7
45.0′–45.5′	50.6	23.0	0.7	2.2	20.8	97.3
45.5′–46.0′	53.7	16.8	0.7	2.4	23.0	96.6
46.0′–46.5′	53.6	17.1	0.6	2.4	22.6	96.3
46.5′–47.0′	54.3	15.7	0.6	2.3	23.2	96.1
47.0′–47.25′	47.7	24.6	0.6	2.2	21.9	97.0
47.25′–47.5′	49.4	25.0	0.6	2.0	19.8	96.8
47.5′–47.8′	46.0	29.8	0.7	2.0	17.9	96.4
47.8′–48.1′	42.9	36.8	0.6	1.9	15.3	97.5
48.1′–48.6′	39.9	43.2	0.6	1.7	13.5	98.9

Analyses and core samples, courtesy of U. S. Bureau of Mines College Park Laboratory.
* Ignition loss expressed as H_2O in fig. 3.

tion cracks were observed in some thin sections of bauxite from Andersonville with 67 per cent Al_2O_3 (Pl. 2, fig. 4). These are connected by a network of open cracks which serve as channels for solution during the alteration of the kaolin to gibbsite. Many of the cracks are partly or wholly filled with gibbsite. Shreds of a mineral with the optical properties of montmorillonite are present in some thin sections.

A color test described by Hendricks and Alexander (1940), in which a blue stain upon treatment with benzidine indicates the presence of clay minerals of the montmorillonite group, was made, taking precautions suggested by Page (1941) to avoid conditions that would interfere with the reliability of the test. The entire core from a depth of 21.5 to 48.6 feet was stained blue after treatment with benzidine, and the deepest blue was in the interval in which the alumina ranged from 41.5 to 53.7 per cent. It seemed desirable to substantiate the presence of montmorillonite in the part of the core that had the highest alumina content. Because the particle size of montmorillonite has been shown to be very small (Ross and Hendricks, 1945), a separation of the particles smaller than 2 microns was expected to contain a concentration of the montmorillonite particles. Four samples of the clay fraction were prepared according to the method of Olmstead *et al.* (1930), in which ammonia is used as the dispersing agent and the time of settling is controlled to obtain the particles less than 2 microns in diameter. The percentage of each

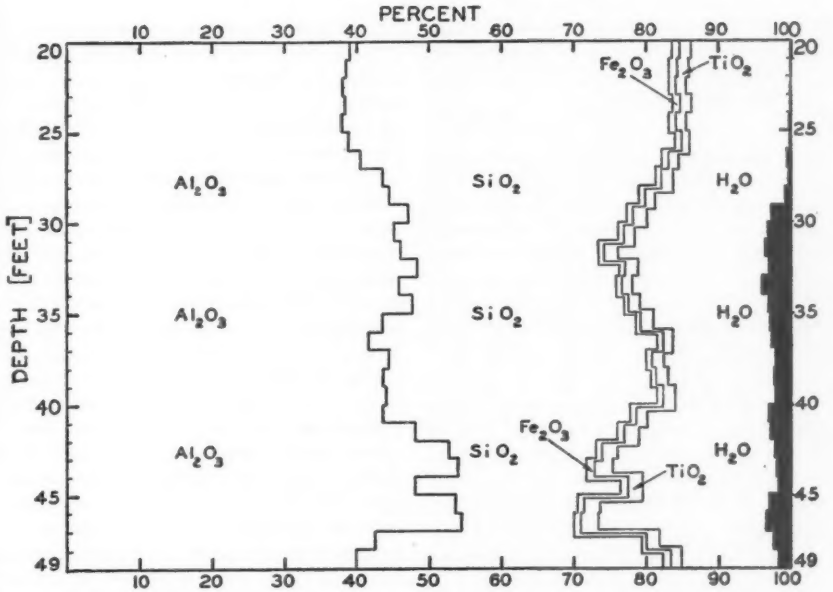

FIGURE 3.—VARIATIONS IN CHEMICAL COMPOSITION WITH DEPTH IN A BAUXITE DEPOSIT IN GEORGIA DERIVED FROM SEDIMENTARY KAOLIN

Analyses by U. S. Bureau of Mines. Solid black represents per cent undetermined, so that each analysis as plotted totals 100 per cent.

TABLE 5.—DATA ON CLAY FRACTION SMALLER THAN 2 MICRONS FROM DRILL HOLE ANG. 18, ANDERSONVILLE DISTRICT, GEORGIA

	1	2	3	4
Interval in feet	21.5–25.5	27.5–36.0	40.0–46.0	46.0–48.1
Fraction of sample in per cent	63.5	63	35.6	29.0
Color with benzidine	blue	blue	blue	blue
X-ray determination	kaolinite	kaolinite gibbsite X-mineral	kaolinite gibbsite X-mineral	kaolinite gibbsite X-mineral
Differential thermal method	kaolinite X-mineral	kaolinite gibbsite	kaolinite gibbsite	kaolinite gibbsite
Petrographic determination	kaolinite muscovite (tr.)	kaolinite gibbsite muscovite (tr.)	kaolinite gibbsite muscovite (tr.)	kaolinite gibbsite muscovite (tr.)
Chemical analysis (see Table 6)	kaolinite +?		kaolinite gibbsite +?	

sample less than 2 microns separated at each interval is given in Table 5, along with the results of x-ray, differential-thermal, petrographic, and chemical examinations. J. M. Axelrod reported that no member of the montmorillonite group was indicated by the x-ray pat-

TABLE 6.—CHEMICAL ANALYSES OF CLAY FRACTION FROM DRILL HOLE ANG. 18, ANDERSONVILLE DISTRICT, GEORGIA, XANTHITANE, HYDROTITANITE AND LEUXOCENE

	1	2	3	4	5	6
SiO_2	40.24	31.23	28.95	1.78	0.32	9.16
Al_2O_3	36.81	43.42	26.19	10.43	22.32	2.72
Fe_2O_3	1.03	1.10	0.78	6.21	5.14	3.07
TiO_2	1.58	2.42	34.43	63.50	53.52	48.10
Cb_2O_5					10.38	
Ta_2O_5						
CaO	none	none	0.06	0.18	3.65	17.44
MgO	0.41	none	tr.	tr.	0.12	0.10
K_2O	0.19	0.32	nd.	nd.	0.17	none
Na_2O	0.34	0.75	nd.	nd.	0.52	none
P_2O_5				4.01		12.81
H_2O-	0.60	0.66	0.36	4.85	0.60	2.04
H_2O+	18.66	20.38	9.12	9.30	3.89	5.14
	99.86	100.28	99.89	100.26	99.82	100.58

1. Clay fraction less than 2 microns in diameter separated from clay at depths of 21.5 to 25.5 ft.; sampled clay overlies bauxite in U. S. Bureau of Mines drill hole Ang. 18. Andersonville, Georgia. Analyst, Samuel H. Cress, U. S. Geological Survey.
2. Fraction less than 2 microns in diameter separated from bauxite at depths of 40.0 to 46.0 ft. in U. S. Bureau of Mines drill hole Ang. 18. Andersonville, Georgia. Analyst, Samuel H. Cress, U. S. Geological Survey.
3. Brown titanium mineral from clay at Ideal mine, 5 miles Northwest of Andersonville, Georgia. Analyst, Rollin E. Stevens, U. S. Geological Survey.
4. Xanthitane, Henderson County, North Carolina. Sample furnished by U. S. National Museum. Analyst, Rollin E. Stevens, U. S. Geological Survey.
5. Hydrotitanite, Magnet Cove, Arkansas. Sample furnished by U. S. National Museum. Analyst, Joseph J. Fahey, U. S. Geological Survey.
6. Leucoxene, in a small pocket associated with ilmenite. Roseland, Virginia. Analyst, Joseph J. Fahey, U. S. Geological Survey.

terns of the four separations which were stained blue with benzidine. Fraction No. 1 (Table 5), separated from kaolin with 38.3 to 38.8 per cent alumina, gave the x-ray pattern of kaolinite. Fractions No. 2, 3, and 4, separated from the portion of the core with 42.9 to 53.7 per cent alumina, gave an x-ray pattern of kaolinite, gibbsite, and lines belonging to an unknown mineral. George T. Faust made differential-thermal curves and reported that No. 1 was kaolinite, and Nos. 2, 3, 4 were kaolinite and gibbsite, but he could not establish the presence of a mineral of the montmorillonite group in the samples. Analyses of separations Nos. 1 and 3 (Table 5), by S. H. Cress of the U. S. Geological Survey, listed as Nos. 1 and 2 in Table 6, do not prove the presence of montmorillonite, nor do they suggest the nature of the unknown mineral reported by Axelrod. Small flakes of muscovite present in the thin sections of these samples would account for the potash and possibly some of the soda and magnesia in these analyses. The conclusions based on these determinations are that kaolinite is the chief clay mineral in the clay, that the high-alumina content of the bauxite is to be attributed to gibbsite, and that montmorillonite is either lacking or is present in amounts too small to be detected by the optical, x-ray, and differential-thermal methods employed. Examination with the electron microscope by the Georgia Kaolin Company and by Davis *et al.* (1950) indicates that some sedimentary kaolins in the Andersonville district, Georgia, and the Aiken district, South Carolina, contain some halloysite; but the methods used in this investigation failed to reveal the presence of any halloysite in the samples studied.

TITANIUM MINERALS IN BAUXITE: The nature of the mineral or minerals in bauxite responsible for titania present is of interest; a few grains of rutile, titaniferous magnetite, or ilmenite can be seen in some thin sections, but the amounts do not appear sufficient to account for all the titania in the bauxite. Irregular brown patches, opaque by reflected light and with the properties of leucoxene, are present in many samples of kaolin and bauxite. Several of these, including one in a sample from the Ideal mine, Andersonville bauxite district, Georgia, were isolated; x-ray patterns by J. M. Axelrod indicate anatase as the titanium mineral and kaolinite as the chief impurity. A sufficient amount of the brown material from the Ideal mine was separated and analyzed. The Al_2O_3 of this sample (No. 3, Table 6) combined with enough SiO_2 and H_2O to form kaolinite utilizes all the SiO_2 and nearly all the water. The slight excess of H_2O might be absorbed by the fine kaolinite particles or by the fine particles of leucoxene. The chemical and x-ray data are in agreement that this leucoxene appears to

have the chemical composition and x-ray structure of anatase.

The writer became interested in the nature of leucoxene and other titanium minerals in connection with the study of diaspore clays of Missouri. Concentrates of leucoxene were isolated and chemically analyzed; Paul F. Kerr reported that the samples gave no x-ray patterns (Allen, 1935a, p. 11–13). These observations indicate that these samples of leucoxene are composed of a hydrous titanium oxide, as suggested by Coil (1933) for the leucoxene in the sandstones of Permian age of Oklahoma. Samples of clay from the Hawaiian Islands derived from basaltic rocks contain 14.06 to 15.28 per cent TiO_2 (Wentworth *et al.*, 1940). W. F. Bradley of the State Geological Survey of Illinois made x-ray patterns of concentrates of leucoxene prepared by the writer and reported that leucoxene coating ilmenite from Norway, a leucoxene concentrate from the clay derived from lava of the Hawaiian Islands, hydrotitanite from Arkansas and xanthitane from Henderson County, North Carolina, have x-ray patterns identical with those of natural and synthetic anatase. Three samples of leucoxene collected by the writer from Roseland, Virginia, in 1941 were found by Gruner to give the x-ray pattern of sphene and anatase or rutile. Also, the xanthitane (No. 4, Table 6) sample gave the x-ray pattern of anatase and a little sphene, and the hydrotitanite (No. 5, Table 6) gave the x-ray pattern of anatase. In thin section the hydrotitanite sample resembles leucoxene, being opaque and gray by reflected light, but xanthitane is yellow and shows high interference colors. New analyses of xanthitane and hydrotitanite were made on materials from the type localities furnished by the U. S. National Museum to be certain that the x-ray patterns were made on samples with the same chemical composition as the type materials. The analysis of xanthitane (No. 4, Table 6) is similar to the original (Shepard, 1856) and is of interest because 0.18 per cent CaO present indicates that the lines representing sphene in its x-ray pattern came from about 1 per cent of sphene in the sample. Independently, Frondel (Palache *et al.*, 1944, p. 588) found that powders of xanthitane and hydrotitanite gave x-ray patterns identical with anatase patterns. The chemical analysis of hydrotitanite from the type locality of Magnet Cove, Arkansas (No. 5, Table 6), furnished by the U. S. National Museum differs from the original (Koenig, 1876) in that it contains 22.32 per cent Al_2O_3 and 10.38 per cent $(Cb, Ta)_2O_5$ and shows that considerable impurities can be present in a sample having the optical properties of leucoxene and its x-ray pattern reveals only anatase. The x-ray pattern of the leucoxene from Roseland, Virginia (No. 6, Table 6), by J. M. Axelrod, indicates anatase, sphene, and apatite in the sample. Ross (Ross and Hendricks, 1945, p. 36) found that leucoxene samples from Roseland, Virginia gave the x-ray pattern of sphene and those from Magnet Cove, Arkansas, gave the x-ray pattern of anatase. Frederickson (1948) stated that his x-ray patterns showed brookite and small amounts of anatase and sphene in a drill core of bauxite from Arkansas in addition to ilmenite, and leucoxene was present in the form of brookite.

The writer's conclusions concerning the titania in bauxite and diaspore clays are: (1) Much of the titanium is present as ilmenite and as the compound known to petrographers as leucoxene. (2) The composition of leucoxene is variable chemically and mineralogically. (3) X-ray patterns of leucoxene are similar to those of anatase, sphene, brookite, and rutile. (4) A small amount of a well-crystallized mineral may show in an x-ray pattern, so it is not known if the x-ray patterns represent most of the leucoxene sample or a few well-crystallized grains. (5) A small amount of water is present in many analyses of leucoxene, but it is not established whether this water is combined as a hydrate of titanium oxide or is absorbed on the smallest particles of an oxide of titanium. (6) Finally, the writer agrees with Tyler and Marsden (1938) that leucoxene should be retained as a petrographic term for the microscopic form in which the titania occurs in rocks and in many clay, bauxite, and laterite samples, but it should not be used as a mineral name.

SIDERITE AND PYRITE: Iron is present in bauxite as limonite or hematite which stains the samples yellow or red, also as ilmenite, siderite, and locally a little pyrite. The siderite occurs as nodules 0.3 to 1.5 mm in diameter or as masses or lenses several inches thick. Many of the

nodules are shotlike forms about 1 mm in diameter. They are composed of nearly pure siderite, for the index of refraction ω is greater than 1.84. Some nodules show a radial structure and have grown after the deposition of the clay, since the radial fibers made room for themselves at the expense of the clay minerals. Pyrite is also later than the clay minerals, and in some cores from the Andersonville district the pyrite surrounds pisolites but only enters them along a few cracks. Some of the pyrite is present as pyritohedrons, a few of which have incomplete outlines, indicating incomplete growth. It is not certain whether they crystallized in the basin of sedimentation or in the clays, after uplift, through the agency of ground-water solutions. In some thin sections of the cores from the Andersonville district, pyrite and siderite occur in adjoining areas. J. M. Axelrod identified chamosite 15 (Fe, Mg)O, $5Al_2O_3 \cdot 11SiO_2 \cdot 16H_2O$ in the x-ray pattern of a bauxite sample collected by C. S. Ross and submitted by C. Milton from Smoky Top (sec. 20 and 29, T. 8 S., E. 1 E.), Pontotoc County, Mississippi; and the writer observed a green mineral having the optical properties of chamosite, in association with siderite in a thin section of the same specimen.

South Carolina, New Jersey, Mississippi, Virginia.—The sedimentary kaolins of the Tuscaloosa formation crop out in South Carolina. In the Aiken district (Fig. 1, loc. 11) pisolites are composed of kaolinite plates forming wormlike books (Pl. 2, fig. 1). Most of the pisolites contain no gibbsite, but in one sample from the Paragon mine John W. Gruner found faint lines belonging to gibbsite along with distinct lines of kaolinite in the x-ray pattern. In thin section, a small area which appears to be gibbsite is surrounded by well-crystallized kaolinite.

Gruner also reported gibbsite in a sample the writer collected from the bottom layer exposed in the open pit at the Hampton Cutter clay mine, Woodbridge, N. J. (Fig. 1, loc. 15). The amount of gibbsite is small and cannot be identified by optical methods. Clays of Cretaceous age with more alumina than the theoretical amount required by kaolinite have been reported from New Jersey (Ries *et al.*, 1904). Edwards (1914) calculated the amount of aluminum hydrates in these clays from the published analyses, but this is the first direct mineralogical evidence known to the writer of the presence of gibbsite in these clays.

Numerous thin sections of the bauxite and associated clays from pits and drill holes in Pontotoc and Union counties, Mississippi (Fig. 1, loc. 14) and Spottswood, Augusta County, Virginia (Fig. 1, loc. 16), show that gibbsite and kaolinite are the chief minerals and that the structures are similar to those described for the deposits of Georgia and Alabama. The geology of the bauxite deposits of Mississippi has been described by Burchard (1925) and by Morse (1923). The bauxite deposits of Mississippi are associated with sedimentary kaolin which rests on the Porters Creek clay of the Midway group of Paleocene age. Many of the so-called pisolites in the bauxite and kaolin are pellets derived from the Porters Creek clay. Glauconite, muscovite, and quartz are present in the pellets; some pellets contain more quartz grains than are present in the surrounding clay (Pl. 2, fig. 2). Veins filled with gibbsite cut the edges of the pellets, and vertically connecting channels filled with gibbsite indicate local downward movement of gibbsite after the deposition of the clay pellets. The beds of kaolin and bauxite are overlain unconformably by the Wilcox formation of Eocene age. Deep valleys mark the unconformity, and erosion channels filled with clay, bauxite, and sands cut the bauxite horizon. Siderite nodules and lenses occur below the bauxite horizon.

Washington, Oregon, California.—Sedimentary clays in Washington, Oregon, and California contain gibbsite in relationships that indicate the alteration of the clay minerals to gibbsite. Near Castle Rock, Cowlitz County, Washington (Fig. 1, loc. 17), gibbsite occurs around the boundaries of pellets and sand-size grains of kaolinite that are surrounded by organic materials outlining bedding, which is parallel to the associated lignite. As the gibbsite occurs around the boundaries of the grains and cuts the grains, it was formed from the clay minerals after they were deposited as fragments by stream action. Evidently the space around the clay pellets and grains provided openings along which ground water carried organic acids that leached silica and deposited gibbsite (Pl.

2, fig. 5). Some specimens contain nearly a third gibbsite and, according to the analyses of the U. S. Bureau of Mines, contain about 42 per cent available alumina.[2] Overlying the kaolinitic clays at the Castle Rock open pit are clays high in iron that are composed of nontronite and beidellite with gibbsite developed around the edges of original grains (Pl. 2, fig. 6). These relations indicate that gibbsite can form from a clay low in alumina when open structures are present in the clay to permit diffusion of ground-water solutions which remove silica and concentrate the alumina in the altered clay (Allen, 1946a, p. 127). Siderite is present in some of the clays either as lenses several inches thick or as small shotlike nodules.

The clays of the Molalla deposit, Clackamas County (Fig. 1, loc. 18), about 30 miles south of Portland, Oregon, are composed chiefly of halloysite-kaolinite with small amounts of montmorillonite, beidellite-nontronite, gibbsite, quartz, feldspars, celadonite, siderite, limonite, zircon, and other minerals. In the lower clays, probably of Miocene age, weathered gravels and volcanic sands and clays containing diatoms represent several types of clays which were deposited by mixing of materials from several sources under fluvial conditions. Apparently some grains were already altered when they were deposited, because in the same thin section some grains are composed of montmorillonite retaining the structure of volcanic glass, and adjacent grains are well-formed books of kaolinite. Also, sand-size fragments with fresh plagioclase are present in the same field of view with sharply bounded fragments that are completely altered to gibbsite. According to analyses of the U. S. Bureau of Mines, one deposit of weathered gravel maintains an available alumina content between 30 and 33 per cent throughout a thickness of 24 feet. Kaolin minerals predominate in the clay types highest in available alumina, and some contain gibbsite that was formed from sand-size grains after their deposition. Siderite nodules and lenses occur at several horizons. Nichols (1944) recognized three zones of weathering within the lower clays that he believed represent three distinct profiles of weathering. Similar zones of alteration in which gibbsite is present also occur in the clay deposit of Castle Rock, where no break in the sedimentary clay sequence is apparent in the field. Two explanations of the gibbsite zones in these clays seem possible. Surface weathering may have altered some grains to gibbsite during an interruption in sedimentation, and later deposition was resumed without leaving a trace of the time interval involved in the weathering. Or, open structures in the clay may have provided somewhat permeable zones along which ground water diffused and carried organic acids that leached silica and later formed gibbsite. Either explanation indicates at least two periods of weathering at Castle Rock and three at Molalla.

In the Alberhill district, Riverside County, California (Fig. 1, loc. 19), pisolitic clays are part of a sedimentary formation containing lignites of Paleocene age. At the pits of the Los Angeles Brick and Clay Company in this district, the pisolitic bone clays contain as much as 46.52 per cent alumina (Dietrich, 1928, p. 354), and others have an alumina content above the 39.5 per cent required by the formula of kaolinite. One sample the writer collected from this locality gave the x-ray pattern of gibbsite and kaolinite, and the thin section of it showed gibbsite developed from the kaolinite of the pisolite. Stauffer (1945) prepared a preliminary map of the clay region of the Santa Ana Mountains with a brief text and a list of the known deposits and of localities possibly worth prospecting.

Replacement by Boehmite

Boehmite was named by De Lapparent (1927) after J. Böhm, the German chemist who first recognized this monohydrate of alumina. Böhm (1925) published a list of localities from which he had observed specimens that contain a monohydrate of alumina with an x-ray pattern different from that of diaspore. This list gave as the only locality of boehmite in the United States "The Linwood-Barton district, U. S. A.,"

[2] According to the U. S. Bureau of Mines, available alumina is the percentage of alumina that can be extracted with 20 per cent H_2SO_4 from the clay after heating it to 700°C. It is based on the weight of the clay dried at 130°C. A clay with more than 20 per cent available alumina is considered a potential high-alumina clay ore (Skinner and Kelly, 1949).

which was entered in the Seventh Edition of Dana's *System of Mineralogy* (1944) as the "Linwood-Barton district, Georgia." No place in Georgia called "Barton" could be located, but there is one in Alabama and a county in Georgia named "Bartow." More than 200 samples of bauxite and clay from Alabama and Georgia, including some from Linwood (Halls on many maps), Bartow County, Georgia (Fig. 1, loc. 20), were examined by x-ray and optical methods; these contained gibbsite but no boehmite. These determinations indicate that gibbsite and not boehmite accounts for the high-alumina content of the bauxite of Georgia and Alabama. An analysis of bauxite given by Hayes (1895) with 75.03 per cent Al_2O_3 from the Armington or Thurman bank in Walker County, Georgia (Fig. 1, loc. 21), may represent boehmite because the alumina content is considerably higher than that required by the formula of gibbsite (Al_2O_3 65.4; H_2O 34.6 per cent). This locality in Walker County is several miles from Linwood, Georgia, but it may be the place from which was obtained the specimen that Böhm determined as the mineral now called boehmite.

The existence of several reported occurrences of boehmite was confirmed during this investigation. The writer visited the boehmite localities reported by McQueen (1943, p. 158) near Swiss and Stolpe, Gasconade County, Missouri (Fig. 1, loc. 22). Specimens were collected and verified as boehmite by x-ray and optical methods. An analysis of one specimen made by the U. S. Bureau of Mines is similar to one made by R. T. Rolufs of the Missouri Geological Survey (Nos. 1 and 2, Table 7). Study of thin sections shows that the boehmite contains a few small oölites of diaspore (Pl. 3, fig. 3). It occurs as lenses interbedded with flint and diaspore clays of Pennsylvanian age. Both boehmite and diaspore replace structures similar to those in the associated flint clay, which is composed entirely of kaolinite. Therefore, it appears that boehmite and diaspore were developed in the flint clay after its deposition by solutions which circulated along the permeable zones at the margins of the oölites.

The high alumina content of the oölites in the flint clay at the Burnt House mine, Olive Hill, Carter County, Kentucky (fig. 1, loc. 24), has been attributed to gibbsite (Galpin, 1912), but x-ray and optical studies made during this investigation indicate that boehmite and not

TABLE 7.—ANALYSES OF CLAYS CONTAINING BOEHMITE

	1	2	3	4	5	6
SiO_2	7.64	4.5	34.76	24.50	6.70	
Al_2O_3	72.12	75.4	48.50	55.00	70.72	85.00
Fe_2O_3	1.65	1.0	1.26	3.40	2.62	
TiO_2	3.06	4.1		2.80	2.66	
CaO	0.22		0.76			
MgO	0.09		0.11			
Na_2O	0.13					
K_2O	0.97					
Others	0.19				0.90	
Ignition loss	14.33	14.3	14.08	15.5	16.40	15.00
Mean Index of Refraction	1.645	1.645	1.635	1.62	1.62	

1. Boehmite clay containing about 2% diaspore; from Swiss (sec. 15, T. 44 N., R. 5 W,) Gasconade County, Missouri. R. T. Rolufs, analyst. Courtesy Missouri Geological Survey.
2. Boehmite clay from same locality as 1. Analysis courtesy of U. S. Bureau of Mines.
3. Boehmite-kaolinite clay, Burnt House mine, Olive Hill, Carter County, Kentucky. A. F. Greaves-Walker, analyst (1907).
4. Boehmite-kaolinite clay, Durham, King County, Washington. Analysis courtesy of U. S. Bureau of Mines.
5. Boehmite pisolites containing some gibbsite and kaolinite, E. H. Miller, analyst. Courtesy of Los Angeles Brick and Clay Co.
6. Theoretical composition of boehmite ($Al_2O_3 \cdot H_2O$).

gibbsite is present. The boehmite consists of definite crystalline aggregates, but the analysis by Greaves-Walker (1907) (No. 3, Table 7) indicates that some kaolinite is probably present. This is not recorded in the x-ray pattern, but if the kaolinite is mixed as finely divided particles it would probably account for the slightly lower mean index of refraction of this material. The boehmite replaces oölites similar to those composed entirely of kaolinite that forms the flint clays of Pennsylvanian age of the Olive Hill district, Kentucky (Pl. 3, figs. 1, 2).

Shales of the Puget group of Eocene age in King County, Washington, contain carbonaceous streaks that locally are replaced by boehmite (Pl. 4, figs. 2, 4). At Durham and Kummer, Washington (Fig. 1, loc. 25), the boehmite was formed in the shales after the deposition of the carbonaceous streaks and at the expense of kaolinite, because kaolinite books are replaced by boehmite (Pl. 4, fig. 1). Also, tiny prismatic and six-sided crystals of boehmite are present locally in the shales of the Puget group (Pl. 4, fig. 5). Some of these are broken, and the separated parts could not have been transported after they were disjoined without pieces being lost. Therefore, these crystals grew in place, probably as ground-water solutions altered the kaolinitic shales to boehmite. Local removal of silica is indicated by the presence of openings and shrinkage cracks in the area high in boehmite. The Durham deposit was explored jointly by the U. S. Bureau of Mines and the U. S. Geological Survey (Nichols, 1945). An analysis of the high-alumina shale at Durham, Washington (No. 4, Table 7) is in agreement with the optical and x-ray evidence that it contains kaolinite and siderite in addition to boehmite. All the iron in this sample was determined and recorded as Fe_2O_3, and CO_2 is included in the ignition loss. Siderite is present in the shales of the Puget group as shotlike nodules and as lenses several inches thick.

E. H. Miller separated the hard pisolites that occur in association with clays of the Eocene age at Alberhill, Riverside County, Calif., and his analysis (No. 5, Table 7) led him to believe they were composed of diaspore. Optical and x-ray determinations on pisolites collected by the writer from this locality indicate that boehmite and gibbsite are the chief minerals (Pl. 3, fig. 5). The ignition loss shows that only a small amount of gibbsite can be present, and the silica content calculated as kaolinite limits the amount of kaolinite present to a few per cent.

Boehmite replaces nodules and is associated with diaspore in the upper part of the iron deposit at Cle Elum, Kittitas County, Washington (Fig. 1, loc. 26; Pl. 3, fig. 4). The geology of this deposit has been described by Zapffe (1944), Lupher (1944), and Broughton (1944).

Most of the black pisolites in the bauxite of lower Eocene age at the Townsend 40 property (SE¼ sec. 11, T. 2 S., R. 14 W.) Saline County and near Berger (NE¼ sec. 9, T. 1 S., R. 12 W.), Pulaski County, Arkansas (Fig. 1, loc. 6) are composed only of gibbsite. However, some contain boehmite associated with gibbsite (Pl. 4, fig. 3), according to x-ray analyses made by Gruner in 1941 and by J. M. Axelrod in 1942 of samples collected by the writer.

The purest boehmite observed is in the material from Swiss, Missouri. Some hand specimens from this locality contain 72 to 75 per cent Al_2O_3 as compared with the theoretical composition of boehmite, 85 per cent Al_2O_3 and 15 per cent H_2O. Optical examination reveals 1 or 2 per cent diaspore; however, the x-ray patterns show no trace of this material but match perfectly those obtained from boehmite from the type locality. Silica and titania are the chief impurities, but optical determinations, x-ray patterns, and differential thermal curves fail to establish how these oxides are combined. It is believed they are present as kaolinite and leucoxene. This boehmite is too fine-grained to permit determination of all its optical constants, but its mean index of refraction of 1.645 agrees with the published data. The slightly lower mean index of refraction of Nos. 3, 4, 5, in Table 7 is probably due to finely divided kaolinite mixed with boehmite. The x-ray pattern of boehmite is definite, and thus it is established that boehmite is an essential component in these clays.

At the six localities where boehmite was studied during this investigation, boehmite is closely associated with gibbsite, diaspore, and kaolinite, indicating a genetic connection between them.

Replacement by Diaspore

In Missouri flint clays composed chiefly of kaolinite were observed (Allen, 1935a) which were changed progressively to diaspore along fracture zones. The analyses in Table 8 represent all transitions from a typical smooth flint clay through rough or oölitic types called "burley clay" to diaspore clay. The term "burley

TABLE 8.—ANALYSES OF FLINT, BURLEY, AND DIASPORE CLAYS OF MISSOURI

	1	2	3	4	5	6	7	8	9
SiO_2	46.5	44.80	33.41	23.39	22.60	9.28	5.46	2.46	
Al_2O_3	39.5	38.84	48.51	55.87	59.32	69.76	74.33	81.29	85.00
Fe_2O_3		0.36	0.65	0.61	1.14	1.14	1.12	.69	
TiO_2		2.44	2.50	3.58	2.14	3.78	3.81	1.95	
CaO		0.05	0.22	0.20	0.18	0.40	0.46	0.20	
MgO		0.10	0.08	0.10	0.22	0.15	0.22	none	
Na_2O		0.30	nd.	nd.	0.66	0.40	nd.	nd.	
K_2O		0.23	nd.	nd.	1.45	0.95	nd.	nd.	
Ignition loss	14.00	13.62	13.78	13.90	12.35	13.37	14.18	14.74	15.00

1. Theoretical composition of kaolinite $Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$.
2. White flint clay, Forbes pit, NW¼ sec. 35, T. 38 N., R. 8 W., Phelps County, Missouri, R. T. Rolufs, analyst.
3. Burley-flint clay, Travis pit, SE¼ sec. 29, T. 41 N., R. 7 W., Maries County, Missouri, H. W. Mundt, analyst.
4. Burley clay, Travis pit, SE¼ sec. 29, T. 41 N., R. 7 W., Maries County, Missouri, H. W. Mundt, analyst.
5. Burley clay, Forbes pit, NW¼ sec. 35, T. 38 N., R. 8 W., Phelps County, Missouri, R. T. Rolufs, analyst.
6. Diaspore clay, Forbes pit, NW¼ sec. 35, T. 38 N., R. 8 W., Phelps County, Missouri, R. T. Rolufs, analyst.
7. Diaspore clay, Holt pit, NE¼ sec. 30, T. 40 N., R. 5 W., Crawford County, Missouri, H. W. Mundt, analyst.
8. Coarse separate from diaspore clay, Holt pit, NE¼ sec. 30, T. 40 N., R. 5 W., Crawford County, Missouri. Separated and analyzed by H. W. Mundt.
9. Theoretical composition of diaspore ($Al_2O_3 \cdot H_2O$).

PLATE 5.—PHOTOMICROGRAPHS OF DIASPORE

FIGURE 1. RING (R) OF DIASPORE [THAT WAS] DEPOSITED AROUND A "SPOT" OR NODULE OF CARBONACEOUS CLAY

Later, diaspore fills in center and forms a nodule of diaspore in the clay of the Mercer shale member of the Pottsville formation. J. Boughman Tract, Clearfield County, Pennsylvania.

FIGURE 2. COLLOFORM STRUCTURE ORIGINALLY KAOLINITE REPLACED BY FIRST-GENERATION DIASPORE (DARK) AND CUT BY VEINS (WHITE) OF SECOND-GENERATION DIASPORE

At right are three rings (R) of diaspore with center nearly completely filled with diaspore. Clearfield County, Pennsylvania.

FIGURE 3. ENLARGEMENT OF COLLOFORM STRUCTURE SHOWN IN FIGURE 2 TO SHOW CONCENTRIC BANDING CHARACTERISTIC OF CLAY

Banding now replaced by fine diaspore (dark, D) of first generation and cut by white veins of second-generation diaspore. Clearfield County, Pennsylvania.

FIGURE 4. DARK NODULES (D) REPLACED BY FIRST-GENERATION DIASPORE AND CUT BY VEINS OF SECOND-GENERATION DIASPORE

Clearfield County, Pennsylvania.

FIGURE 5. DIASPORE FILLING VEINS AND REPLACING STRUCTURES IN UPPER PART OF CLE ELUM IRON DEPOSITS

At left center is chromite (C) cut by veins of chlorite which is partly replaced by diaspore near the lower fourth of photograph; this diaspore vein swings to the right and joins main vertical diaspore vein (D). Ovate areas were clay pellets of kaolinite and boehmite now partly or wholly replaced by diaspore. Kittitas County, Washington.

FIGURE 6. DIASPORE CLAY REPLACED BY LIMESTONE

Under high magnification diaspore (D) at center has crystal faces at both ends. This is the only occurrence of doubly terminated crystals of diaspore observed by the writer. Phelps County, Missouri.

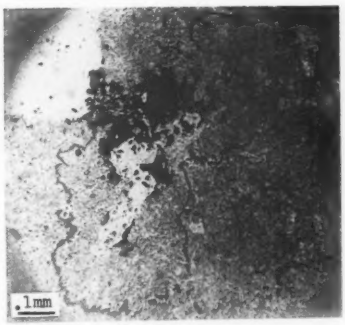

FIGURE 1

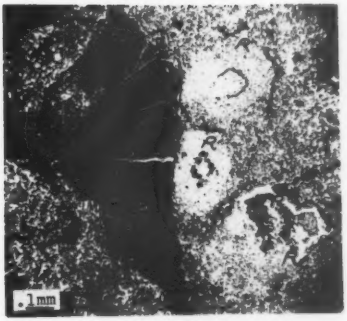

FIGURE 2

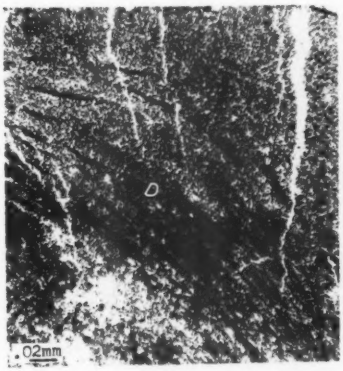

FIGURE 3

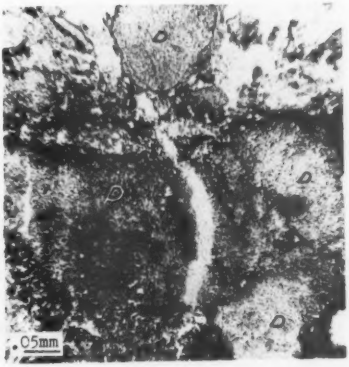

FIGURE 4

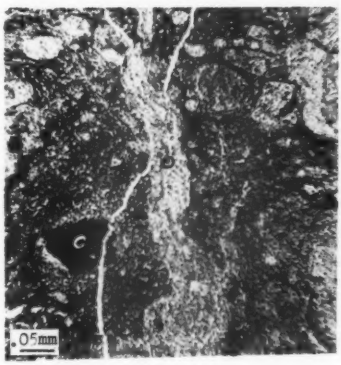

FIGURE 5

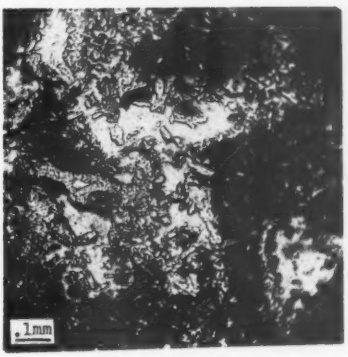

FIGURE 6

PHOTOMICROGRAPHS OF DIASPORE

FIGURE 1

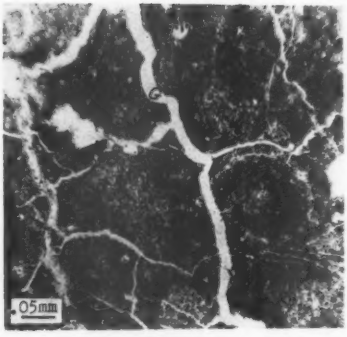

FIGURE 2

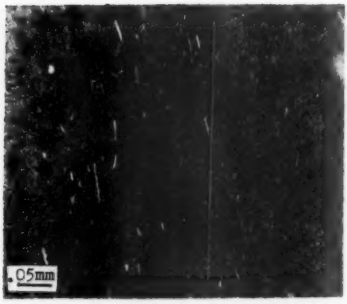

FIGURE 3

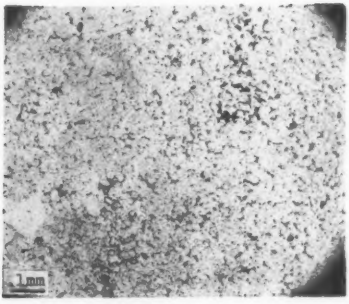

FIGURE 4

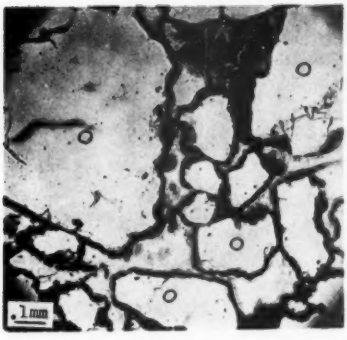

FIGURE 5

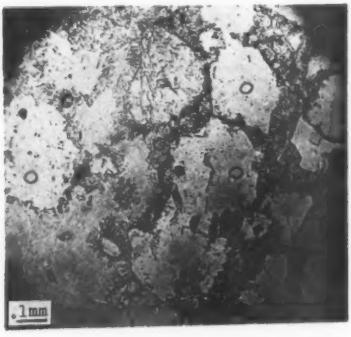

FIGURE 6

PHOTOMICROGRAPHS ILLUSTRATING MIGRATION AND RESILICATION OF GIBBSITE

clay" is used in the diaspore district for a rough-feeling clay containing "burls" or oölites and having an alumina content between 45 and 68 per cent. A diaspore clay contains at least 68 per cent alumina. The roughness of the clay is caused by needlelike crystals of diaspore, which increase in amount with increase of alumina. The composition of the isotropic matrix around the diaspore grains also varies directly with the amount of diaspore, since the index of refraction of the matrix increases directly as the alumina content of the clay. The isotropic matrix does not consist of a single mineral but is a mixture of several minerals of colloidal dimensions. Photomicrographs can be selected (Allen, 1935a, Pl. III) showing the progressive change of oölites in the flint clay composed of kaolinite to those composed entirely of diaspore. The chemical composition of the coarse fraction representing mainly oölites has been shown by Mundt to approach closely that of diaspore (Nos. 8 and 9, Table 8). The determination of the high-alumina mineral of the oölites as diaspore has been verified in concentrates prepared by the writer, by optical determinations, x-ray patterns, and differential thermal curves. The conclusion reached as a result of his investigations is that ground-water solutions carrying carbonic acid along fracture zones removed silica from the kaolinite flint clays to form the diaspore clays.

In the summer of 1942, Dr. John M. Parker, III, visited the high-alumina clay district in Pennsylvania for the U. S. Geological Survey and brought back specimens from the Curwensville and Morgan Run districts in Clearfield County (Fig. 1, loc. 27) that were studied petrographically by the writer. In July, 1943, the writer visited the important diaspore clay districts of Pennsylvania with Richard M. Foose and brought back additional specimens for study. In 1944, petrographic study was extended to the diaspore districts in Clinton County (Fig. 1, loc. 28) in connection with a cooperative project of the U. S. Geological Survey, the U. S. Bureau of Mines, and the Pennsylvania Topographic and Geologic Survey.

In 1942, the writer recognized that diaspore replaced colliform and nodular structures like those present in the flint clays of Pennsylvania, that two generations of diaspore were deposited, and that a series of mineral changes, including the formation of diaspore, siderite, and pyrite, modified the original composition of the clays and increased their alumina and iron content. The spotted clay and the oölites or pellets of kaolinite in the flint clay are replaced by diaspore, but the structure is preserved (Pl. 5, fig. 4). Some spots consist mainly of carbonaceous clay material that is darker than the surrounding matrix clay. Other spots consist of light clear kaolinite surrounded by evenly stained light-brown carbonaceous clay. The replacement is selective, since certain clay areas and spots are surrounded by a ring of diaspore (Pl. 5, figs. 1, 2) that was gradually filled in with diaspore of two generations. Colloform structures in the flint clays consisting of swirls and concentric bands are retained in the nodular

PLATE 6.—PHOTOMICROGRAPHS ILLUSTRATING MIGRATION AND RESILICATION OF GIBBSITE

FIGURE 1. GIBBSITE FILLING VEIN (WHITE, G) CUTTING BAUXITE DERIVED FROM SEDIMENTARY KAOLIN OF TUSCALOOSA FORMATION
Wilkinson County, Georgia.

FIGURE 2. GIBBSITE FILLING IRREGULAR VEIN (WHITE, G) CUTTING EARLY STRUCTURES IN BAUXITE
Andersonville district, Sumter County, Georgia

FIGURE 3. SECTION OF A GIBBSITE TUBE WITH MUSCOVITE FLAKES ARRANGED WITH LONG DIMENSIONS PARALLEL TO THE LENGTH OF TUBE
Toomsboro, Wilkinson County, Georgia

FIGURE 4. NODULE OF SECONDARY GIBBSITE CONSISTING OF INTERLOCKING PLATES OF GIBBSITE WITH NO RELICT STRUCTURE
Dark areas are iron oxide stains. Near Salem, Oregon.

FIGURE 5. CELLULAR KAOLIN WITH THIN WALLS OF KAOLINITE SURROUNDING OPEN CELLS (WHITE, O)
Thin cleavage flakes growing perpendicular to the walls give fibrous appearance to kaolinite. (K) is a "book" of kaolinite. Cowpen Branch, Wilkinson County, Georgia.

FIGURE 6. OPEN CELLS (WHITE, O) SURROUNDED BY GIBBSITE (G)
Cowpen Branch, Wilkinson County, Georgia.

types of diaspore clays (Pl. 5, figs. 2, 3). The second-generation diaspore is colorless and fills veins that cut across the early minerals and structures and the fine-grained diaspore of the lying flint clay containing nodules or spots which are locally replaced by diaspore. The same clay minerals occur between nodules of the flint clay and in the block clay. However,

TABLE 9.—ANALYSES OF SEMI-FLINT, BLOCK, FLINT, AND DIASPORE CLAYS OF PENNSYLVANIA

	1	2	3	4	5	6	7	8	9
SiO_2	46.5	46.76	43.40	43.47		26.76		4.00	
Al_2O_3	39.5	35.13	37.55	38.37	49.94	54.50	61.52	75.72	85.00
Fe_2O_3		1.52	1.06	1.29		.80		2.17	
TiO_2		1.57	1.77	1.72		2.26		2.75	
CaO		0.18	0.13	0.04		0.18		0.10	
MgO		0.73	0.43	0.28		0.30		0.25	
K_2O }		2.60	1.51	0.96		0.95		0.38	
Na_2O }									
Ig. Loss	14.00	11.17	13.40	13.16	13.46	13.50	13.88	14.40	15.00

1. Theoretical composition of kaolinite ($Al_2O_3 \cdot 2SiO_2 \cdot 2H_2O$).
2. Semi-flint clay.
3. Block clay.
4. Flint clay.
5. Nodule-block clay (incomplete analysis).
6. Green nodule clay.
7. Fine-grained nodule clay (incomplete analysis).
8. "Burnt" nodule clay.
9. Theoretical composition of diaspore ($Al_2O_3 \cdot H_2O$).

Analyses courtesy of General Refactories Co. (Foose, 1944, p. 568).

first generation (Pl. 5, figs. 2, 3, 4). Also, it forms a rim of coarsely crystalline diaspore with prominent cleavage around the edges of the nodules replaced by fine-grained diaspore. Siderite is later than the diaspore of both generations. It replaces the clay minerals and the diaspore and fills the space in the center of the nodules made when silica was removed from the clay minerals to form first-generation diaspore. Pyrite appears to be the last mineral deposited, but its time relation to siderite is not clearly shown. Locally, oxides of iron have formed by alteration of siderite and pyrite.

The diaspore clay occurs near the middle of the underclay of the Mercer shale members of the Pottsville formation of Pennsylvanian age (Foose, 1944). No apparent break or disconformity separates it from the overlying block clay, which consists of kaolinite with a little hydromica or muscovite and a few grains of rutile, zircon, and tourmaline. Close to the overlying coal of the Mercer shale member, the clay is carbonaceous. The chemical and mineralogical composition of the block clay at the top is not very different from that of the underthe block clay at the top lacks nodules and has less porosity. The increased porosity around the nodules localized the alteration of the clay minerals to first-generation diaspore. As silica was removed from the clay minerals, open spaces were formed and accelerated the formation of diaspore. At a certain stage, chemical conditions of the ground water within the deposits allowed some diaspore formed at upper levels to be carried downward and deposited as second-generation diaspore. Conditions were later modified, and siderite and pyrite were deposited. The silica leached from the clays was carried outside the area. Downward circulation of the reacting ground-water solutions was hindered by the relatively impermeable semi-flint and plastic clays containing kaolinite, muscovite or "hydromica," and quartz.

Measurement of Porosity

Many of the methods of determining porosity cannot be used on clays because of their plasticity and softening in water. The apparent porosity as expressed by Ries (1927, p. 242) was

selected because it gave a means of comparing the relative porosity of several nonplastic types of clays and bauxite under uniform conditions with results that could be repeated and checked fairly accurately. Apparent porosity is expressed by the volume of the open pores and can be determined from the following formula.

$$\text{Apparent porosity} = \frac{(W - D)\,100}{W - S}$$

W is the weight of the saturated specimen
D is the weight of the dry specimen, and
S is the weight of the saturated specimen suspended in water.

The specimens were saturated by placing them in a beaker of water that was kept in a vacuum for several hours until the weight of the saturated specimen was constant.

The apparent porosity and the approximate alumina of the 15 samples tested are given in Table 10. The conclusions based on these results are:

(1) The apparent porosity of flint clays ranges from less than 1 per cent for noncarbonaceous types to 31 per cent for some carbonaceous types.

(2) In general, the boehmite and diaspore clays are more porous than the flint clays at the same locality. This supports th e petrographic evidence that removal of silica accompanied the formation of diaspore and gibbsite and increased the porosity.

(3) The "burnt" nodule clay (No. 3, Table 10) with 70 per cent Al_2O_3 has only 18 per cent apparent porosity. Evidently, second-generation diaspore reduced the porosity by filling in some of the open pores that were developed when silica was removed to form the first-generation diaspore.

(4) The apparent porosity of a specific type of clay varies directly with alumina content. A boehmite-kaolinite clay with 10 per cent apparent porosity and 48 per cent Al_2O_3 (No. 10, Table 10) was formed from a kaolinite clay like No. 9 having about 7 per cent apparent porosity and about 39 per cent Al_2O_3. This suggests that replacement of the oölites by boehmite (Pl. 3, figs. 1, 2) requires removal of silica, increase of Al_2O_3 and of pore space, and not mere substitution of boehmite for kaolinite. Furthermore, the apparent porosity of the boehmite clays from other localities with 73 per cent Al_2O_3 is higher than that of No. 10 with 48 per cent Al_2O_3. Increase of porosity accompanies increase of Al_2O_3 and reduction in silica that is expressed in the chemical analyses of the specimens. (*See* Table 7.)

(5) Petrographic evidence suggests that the cellular clay from Cowpen Branch, Georgia, was formed from a cellular bauxite with the same structure but not necessarily identical original apparent porosity. The addition of silica to gibbsite to form the kaolinite of the cellular clay (Pl. 6, fig. 5) required a reconstruction or rearrangement of the elements to keep the ap-

TABLE 10.—APPARENT POROSITY AND APPROXIMATE ALUMINA OF SAMPLES TESTED

	Apparent Porosity (per cent)	Approximate Al_2O_3 (per cent)
1. Spotted flint clay, Jeffries mine, Pennsylvania	0.1	38
2. Spotted flint clay below *1*, Jeffries mine, Pennsylvania	1	38
3. "Burnt" nodule clay, W. D. Little mine, Pennsylvania	18	70
4. Black flint clay (carbonaceous), Stolpe, Missouri	31	38
5. Boehmite clay underlies 4, Stolpe, Missouri	28	73
6. Boehmite clay, Swiss, Missouri	23	73
7. Flint clay, Forbes pit, Phelps County, Missouri	1.6	39
8. Diaspore clay, Forbes pit, Phelps County, Missouri	29	70
9. Flint clay, Olive Hill district, Kentucky	7	39
10. Boehmite-kaolinite clay, Olive Hill district, Kentucky	10	48
11. Cellular clay, Cowpen Branch, Wilkinson County, Georgia	35	39
12. Cellular bauxite, Cowpen Branch, Wilkinson County, Georgia	32	57
13. Clay "chimney rock type," Toomsboro, Georgia	28	39
14. Clay over 13, Toomsboro, Georgia	35	39
15. Bauxite, same pit as 13, 14 Toomsboro, Georgia	30	50

parent porosity at about the same amount and still permit the addition of silica to the sample. The same difficulty is present if clay of the "Chimney rock type" (No. 13, Table 10) was formed by resilication of bauxite having the same apparent porosity as the bauxite present in the same pit (No. 15, Table 10). Possibly some gibbsite is liberated during or in advance of resilication and migrates into another part of the deposit.

Migration of Gibbsite and Diaspore

Gibbsite and diaspore are found in secondary relationships which leave no doubt that they can migrate from the places where they were first formed to new situations.

Gibbsite fills secondary veins which cut kaolin and bauxite of the Tuscaloosa formation of Wilkinson County, Georgia (Fig. 1, loc. 9), in a manner indicating that gibbsite is being transferred from the matrix into fractures (Pl. 6, fig. 1).

In the Andersonville district, Sumter County, Georgia (Fig. 1, loc. 12), and in Barbour and Henry counties, Alabama (Fig. 1, loc. 13), pisolitic structure is developed in mixtures of fine-grained gibbsite and kaolinite. Crystalline gibbsite fills cracks that cut across the early structures of the bauxite (Pl. 6, fig. 2) and affords evidence of a secondary movement of gibbsite.

Near Toomsboro, Wilkinson County, Georgia (Fig. 1, loc. 9), cylindrical tubes, one-fourth to three-fourths of an inch in diameter and 2 to 3 inches long, lie exposed on the surface detached from the clay. Optical, x-ray, and differential-thermal studies indicate that the tubes are composed of gibbsite with minor amounts of kaolinite, muscovite, and detrital quartz. A partial chemical analysis gave 59 per cent Al_2O_3 for some of these tubes. Thin sections cut parallel to the long axes of the tubes show residual areas of kaolinite and plates of muscovite arranged with their long dimensions parallel to the length of the tubes (Pl. 6, fig. 3). A similar gibbsite tube containing more detrital quartz grains was found in place in clay pit No. 10 of the Georgia Kaolin Company, Dry Branch, Twiggs County, Georgia (Fig. 1, loc. 8), where the relations in thin section indicate that gibbsite has replaced not only kaolinite but also detrital quartz. These secondary tubelike or stalactitic concentrations of gibbsite were developed by the movement of gibbsite which oriented the muscovite plates parallel to the length of the growing tube.

Near Salem, Oregon (Fig. 1, loc. 29), nodules of gibbsite contain 60 per cent alumina without a suggestion of the structure of the parent rock. In thin section, they consist of interlocking grains of well-crystallized gibbsite (Pl. 6, fig. 4). X-ray patterns indicate only gibbsite, but a differential-thermal curve made by S. S. Goldich shows a pronounced gibbsite peak and a small peak which he estimates to be 5 per cent kaolinite. Other nodules of gibbsite occur in the region, and some show the outlines of the plagioclase laths of the basaltic rocks from which the gibbsite was formed by leaching of silica (Allen, 1948). The small percentage of kaolinite in the nodules suggests it is the remnant of the clay formed during the first phase of weathering of plagioclase to clay. Nodules of crystalline plates of gibbsite without a trace of the structure of the original rock also occur in the bauxite district of Arkansas. These are interpreted by the writer as representing secondary gibbsite that has been concentrated by migration and recrystallization.

Two generations of diaspore occur in the nodular clays of Clearfield and Clinton counties, Pennsylvania (Fig. 1, locs. 27, 28). The first-generation diaspore is fine-grained and was formed by leaching of silica from the clay minerals. It replaces pellets and other structures of the flint clays that were originally composed of kaolinite. The second-generation diaspore is coarse-grained and fills veins that cut across the early fine-grained diaspore (Pl. 5, figs. 2, 3, 4). It forms a rim around centers of fine-grained nodules and was deposited in open spaces provided by the removal of silica from the clay. In the upper part of the Cle Elum iron deposit, Kittitas County, Washington (Fig. 1, loc. 26), diaspore replaces chlorite that is present in veins cutting chromite and other early minerals and structures (Pl. 5, fig. 5). It also fills veins that give no evidence of having originally been occupied by chlorite, and it replaces early minerals and structures, including pellets that were originally kaolinite

and boehmite. Late movement of diaspore along open cracks is indicated at these localities.

Information on the conditions under which the migration of gibbsite and diaspore takes place is not available. Two modes of transportation are suggested: (1) migration in colloidal suspensions; (2) transfer of their constituents in true solution. Evidence regarding the ease with which colloidal suspensions of gibbsite and diaspore can be formed is lacking. The evidence suggests that conditions probably exist under which these minerals can migrate in this manner. In addition, there are indications that the constituents of these hydrous aluminum oxides under some conditions are transferred as true solutions. During the process of desilication, hydrous aluminum silicates are decomposed chemically, and silica is removed in solution. According to Roy (1945) and others, some of the silica in natural waters is carried in true solution, probably ionic. If the constituents of hydrous aluminum oxides are liberated as true solutions during chemical decomposition, it appears likely that they would be transported as such rather than changed over to the colloidal state before deposition in the secondary relationships in which they have been observed. Evidence for the migration of clay minerals has been presented (Allen, 1945); the writer recognizes that the amount of ground water which circulates through a clay under conditions of good drainage determines whether the clay minerals remain in place or are carried away. Whether conditions of good drainage have a similar effect on the hydrous aluminum oxides is open to discussion, but ground-water circulation can probably modify the conditions within the chemical system where solution and deposition occur. Whatever may be the controlling conditions for migration, this process offers the most likely explanation of the high concentration of hydrous aluminum oxides in the secondary relations observed.

MIGRATION OF CLAY MINERALS: The migration of clay minerals has been observed and described at many localities (Allen, 1945). All the common clay minerals—kaolinite, halloysite, dickite, montmorillonite, and nontronite—have been observed filling cracks that cut across early minerals and structures, and occupying vesicles and open cavities in clays and other rocks. The widespread occurrence of these minerals in secondary relationships not only indicates migration, but also that they migrate with ease, probably as colloidal suspensions under conditions of good drainage.

The migration of clay minerals into clay masses modifies the composition of the resulting clay mass. Specimens photographed (Allen, 1945) show several systems of clay-filled veins resulting from successive fracturing and offsetting of veins healed by early clay minerals. The migration of clay minerals into bauxite masses adds silica and decreases the proportion of alumina in the resulting mixture. The filling of cracks and open cavities in a bauxite deposit with clay minerals decreases porosity and permeability, but, where alternate wetting and drying occur, new cracks may form during drying and be filled with clay minerals during the wet periods. The addition of clay minerals to a bauxite deposit increases its volume and destroys or obscures the original structure of the parent materials by filling all the available openings and crowding aside the old material.

VEINS OF KAOLIN MINERALS IN BAUXITE IN ARKANSAS: The veins of kaolin penetrating the bauxite of Arkansas (Goldman and Tracey, 1946) the writer considers to be the result of migration of kaolin minerals rather than the result of resilication of gibbsite. After the bauxite of Arkansas was formed, cracks developed in it at the surface; when deposition of clays of the Wilcox formation commenced, some of the clay minerals were carried into and along the cracks. The migration of clay minerals into cracks and openings has been described at many localities where bauxite is not present (Allen, 1945), and the process is not necessarily connected with the formation of bauxite.

The distortion and destruction of original structures of the parent rock that may accompany the introduction of kaolin minerals (Goldman and Tracey, 1946, p. 569) result from the increase in volume caused by the addition of new material that not only fills all the available openings but also crowds aside and disturbs the old materials.

RESILICATION

Harrison (1933) recognized the formation of crystalline kaolinite from finely divided gibbs-

ite by reaction with silica carried upward by capillary moisture during evaporation in dry seasons which alternate with wet seasons in tropics. In this process, which he called resilication, one portion of the alumina liberated from gibbsite (theoretical 65.4 per cent Al_2O_3) combines with silica in the ratio 39.5 per cent Al_2O_3 to 46.5 per cent SiO_2 to form kaolinite. Another portion of the aluminum trihydrate (theoretically the difference between 65.4 Al_2O_3 in gibbsite and the 39.5 Al_2O_3 required to form kaolinite) is taken into solution and migrates to another part of the deposit to be deposited as films, veins, and masses of gibbsite of spheroidal and other shapes. As a surplus of Al_2O_3 is available under conditions of resilication, the process goes on as long as silica is supplied and the conditions remain favorable for resilication, so no silica is available for precipitation as quartz or opal. Hardy and Rodrigues (1939) and Alexander, Hendricks, and Faust (1941) show that, where silica is being liberated by mineral weathering in close proximity to gibbsite, resilication of gibbsite to form kaolinite may take place.

The most convincing evidence of resilication observed by the writer consists of cellular kaolinite masses from Cowpen Branch, about 2½ miles south of Clayfields, Wilkinson County, Georgia (Fig. 1, loc. 9; Pl. 6, fig. 5). The cells of kaolinite range from 0.2 mm to about 1 mm in their longer diameters, and the walls of kaolinite are 0.02 to 0.03 mm thick. The apparent porosity of a typical specimen is 35 per cent (No. 11, Table 10). This open structure in a clay is unusual, but a cellular structure is common in bauxite and is formed during removal of silica from a clay when the gibbsite is formed. In fact, cellular bauxite occurs at the same locality only a short distance away. A specimen of it has an apparent porosity of 32 per cent (No. 12, Table 10). It has cells about the same size as those of the cellular kaolinite, but some of the walls of gibbsite are about the same thickness as the kaolinite ones and others are 2 or 3 times as thick (Pl. 6, fig. 6). In thin section, the cellular kaolinite contains leucoxene as grains 0.2 to 0.4 mm across and as tiny grains scattered through the low birefringent kaolinite which forms the center of the cell walls (Pl. 6, fig. 5). From the center, fibers or tiny sheets of kaolinite project from the walls, and locally books of kaolinite up to 1 mm long are attached to the walls and conform to their curvature. A few books contain sheets of a mineral that has birefringence between 0.03 and 0.04, parallel extinction, is interlayered with the cleavage sheets of kaolinite, and forms small fans at the edges of the books. It is probably the kaolin mineral of Ross and Kerr (1931), which has been mistaken for muscovite but has the same composition as kaolinite, except for having one molecule of water in place of two. In one part of the thin section of the cellular gibbsite, isotropic and low-birefringent kaolinite surrounds and replaces the fine-grained gibbsite but not the larger crystals which have a length of 0.1 mm.

The writer's explanation of the origin of the cellular structure of the kaolinite is that the structure was formed when the removal of silica resulted in the formation of gibbsite in a clay of the Tuscaloosa formation. Later, ground water carried silica in solution, and kaolinite was formed by the reaction of silica with finely divided gibbsite. Some alumina was removed from the specimen so that its apparent porosity remained about the same throughout the addition of silica. The secondary kaolinite formed by resilication of gibbsite is opaque and gray or white by reflected light. It was hoped this opacity could be used to determine whether a clay was formed by resilication, but the test is inconclusive because some forms of finely divided gibbsite and leucoxene show a similar opacity by reflected light. It was necessary to check each specimen carefully by oil immersion, by x-ray, and by differential thermal methods to establish which of the opaque minerals was present.

Some varieties of clay, called "chimney rock" (Shearer 1917, p. 121), occur in association with bauxite in Wilkinson County, Georgia (Fig. 1, loc. 9). Their opacity, hardness, massiveness, and relationship to bauxite suggest that some "chimney rock" specimens were formed by resilication of gibbsite. The apparent porosity of "chimney rock" (No. 13), the clay over the "chimney rock" (No. 14), and bauxite (No. 15) from the same pit at Toomsboro, Georgia, are given in Table 10.

Areas of dense, gray clay at the Ideal mine,

Andersonville district, Georgia (Fig. 1, loc. 12), at the Ratcliffe mine (NW¼ sec. 24, T. 1 N., R. 12 W.) a mile north of Sweet Home, and other localities in the bauxite district of Arkansas (Fig. 1, loc. 6) were probably formed by resilication of gibbsite.

The conditions under which resilication takes place are no better known today than when Harrison (1933) first described the process. C. S. Ross called the writer's attention to a thin section of bauxite from Mississippi in which gibbsite embays and replaces detrital quartz without resilication of the gibbsite to kaolinite. Similar occurrences have since been observed by the writer in thin sections of clays from Mine No. 10, Georgia Kaolin Company, Dry Branch, Georgia, and from clays in Wilkinson County, Georgia. Apparently, the conditions permitting gibbsite replacement of quartz are not those favorable for resilication. An angular grain of detrital quartz 0.3 mm in diameter occurs in the specimen of cellular kaolinite from Cowpen Branch, Georgia, with a sharp contact and no signs of embayments. Two conditions controlling this change may be a very small size of grain of the gibbsite to be resilicated and a chemical system of favorable composition and concentration. If some silica in natural waters is carried in true solution, probably ionic as Roy (1945) states, such ionic solutions of silica would be more favorable for resilication during liberation of alumina from gibbsite than colloidal suspensions of silica would be.

SILICA IN THE CLAYS OF GEORGIA: Stull and Bole (1926) believe that the hardness of the hard clays of Georgia is due to "free silicic acid" that is lacking in the soft clays. The presence of "silicic acid" was considered the cause of the difficulty in slacking and filter-pressing the hard clays. Spring waters flowing on the top of the clays are said to be opalescent and high in silica. As the possible presence of silica in spring waters and in clays of Georgia is of interest in connection with the desilication and resilication, search was made for evidence of free silica in the clays of Georgia.

At the mine of the Harbison-Walker Mining Company, 2 miles southeast of Gibson, Glascock County, Georgia (Fig. 1, loc. 30), a sample of silica jelly was collected by the writer in 1941. When dried in air for several hours, it has the properties of opal, being isotropic and having an index of refraction of 1.465. An x-ray pattern made by Dr. John W. Gruner showed lines of cristobalite given by many specimens of opal. In 1943, R. M. Thompson collected a quart milk bottle of silica jelly from a cavity exposed by blasting during operation of the mine at the same locality. As long as it retained its original content of water, it could be shaken like jelly, but when it was dried at room temperature, it became hard, broke with a conchoidal fracture, and had an index of refraction of 1.465. Earlier Smith (1929, p. 352) had collected a sample of silica jelly contaminated with fragments of flint clay and plant remains and published the following analysis of it:

	Per cent
SiO_2	72.49
Al_2O_3	8.84
Fe_2O_3	3.22
TiO_2	0.90
CaO	0.00
MgO	tr.
SO_3	0.72
P_2O_5	0.10
Moisture	6.19
Loss on ignition	7.73

In only one other clay from Georgia was collodial silica or opal found. This was a soft clay from pit No. 7 of the Georgia Kaolin Company, Dry Branch, Twiggs County, Georgia (Fig. 1, loc. 8), which showed cristobalite lines in the x-ray pattern made by Gruner. No evidence could be found that hard clays contain free silicic acid and that soft clays do not. The hardness seems to be connected with the size of grain of the kaolin particles. The silica jelly and opal found in both hard and soft clays indicate that silica is carried by ground water in Georgia and is not characteristic of any one type of clay.

Nodules of chalcedony or opal have been observed by Bridge (1949) in the clay pit of the Cold Spring Mining Company near Cold Springs, Augusta County, Virginia. Some of the nodules, which are 3 inches or more in diameter, have a pale milky color and irregular outline and look like hardened masses of colloidal silica.

Free silica occurs in many clays as finely

divided quartz, but it is difficult to establish if this quartz was deposited with the clay minerals in the basin of sedimentation or if it represents colloidal silica carried there later and recrystallized.

Review of Previous Ideas of the Origin of Bauxite, Diaspore, and Boehmite

Original Sediments

Many theories of origin of bauxite, diaspore, and boehmite have been proposed. Some of these have been applied to the deposits included in this investigation and will be examined in the light of the evidence obtained during these studies.

Bauxite, laterite, and diaspore deposits have been regarded as original high-alumina sediments. Mallet (1881) considered that the laterite of India and the iron clays or bauxite of Ireland were lacustrine sediments. He described the bauxite of Ireland as a single horizon which separated the basaltic lavas of Antrim into two series and which was associated with sedimentary rocks containing plant remains of the Miocene epoch indicating a warm or subtropical climate. Spencer (1893) believed that the bauxite of Georgia-Alabama represents the aluminous equivalent of bog iron ore deposited in a lagoon. The possibility was once considered that the diaspore clays of Pennsylvania were deposited as diaspore sediments because no break could be found between them and the associated clays. This idea does not solve the problem of origin but merely transfers it to the source of the sediments. Also, it will not explain the replacement of the primary structures of these sediments and the series of alterations indicated by the chemical and petrographic evidence.

Hot Springs

Hayes (1895) proposed that the bauxite of Arkansas was formed by strong alkaline or saline waters acting on the feldspathic components of the heated syenite. He (1894) also observed gibbsite associated with the bauxite of Georgia and believed it was formed by heated ascending waters, which decomposed pyrite in the underlying shale, brought aluminous solutions in contact with carbonate of lime, and precipitated bauxite near the surface. Cross (1896) reported a rhyolite at Rosita Hills, Colorado, to be altered to diaspore, quartz, and alunite. The origin of bauxite by the action of hot springs was applied to the bauxite deposits of France by Coquand (1870) and by Auge (1888), but this theory has lost supporters in recent years. Where the writer has made detailed mineralogical studies, hydrothermal and solfataric minerals are absent, and field evidence of hot springs action is lacking.

Sulfuric Acid

Shearer (1917, p. 125) offered the hypothesis that the bauxite of Georgia was due to the action of hydrogen sulfide introduced into the lagoons in which kaolin was being deposited. The clay beds underlying the kaolin of the Tuscaloosa formation are reported by him to contain finely divided pyrite, which was the source of the hydrogen sulfide solutions. Shearer believed that descending waters containing oxygen acted on pyrite, produced ferrous and ferric sulfates, and oxidized the remaining sulfur to sulfuric acid. On penetrating below the zone of oxidation, this free sulfuric acid attacked other sulfides and liberated hydrogen sulfide. The hydrogen sulfide solutions formed groups of sulfur springs. These hydrogen sulfide springs contributed sulfuric acid to the bodies of water in which kaolin was present, and aluminum sulfate was obtained by the decomposition of the kaolin. The aluminum sulfate was carried to a position where hydrogen sulfide was present and the waters were neutral or alkaline. The aluminum sulfate was hydrolyzed and precipitated as aluminum hydrate, and hydrogen sulfide and sulfuric acid were regenerated.

The pyrite the writer observed in the cores of bauxite and clay of Georgia is not oxidized and shows no evidence of decomposition to supply sulfuric acid for the formation of bauxite. Also, experiments show that sulfuric acid leaches alumina and iron from a clay, leaving it higher in silica than it was originally (Allen, 1935a, p. 22–24). The petrographic and chemical evidence indicates that silica was removed from the kaolin of Georgia during the early

stages of the formation of bauxite to form the open spaces and textures, which accounts for high apparent porosity.

Organisms and Organic Acids

Holland (1903) advanced the idea that some lowly organisms can separate alumina from silica which is necessary for their life processes. The unused alumina is soluble and is removed by alkaline solutions. Thiel (1927) has shown that microorganisms aid in liberating aluminum, but the silica of the clay acted on by bacteria is higher than it was before alumina was removed. Bacteria would not aid in desilication of clay in place, but they would liberate alumina and make it available for enrichment through migration.

Behre (1932) directed attention to the lignite which overlies some bauxite deposits in Arkansas. He cited the opinions of German investigators that the decomposition of the underlying rocks to clay or bauxite is accomplished by humic and lactic acids in the swamp waters while the vegetation was accumulating and decaying. Behre suggested that alteration by these reagents could be in progress in the bauxite district of Arkansas today. Fetzer (1946) studied the solvent action of humic acids and true organic acids and concluded that humic acids are lyophilic colloids that are insoluble in water and impart no acidity to the water. The solvent action of humic acids is due mainly to carbon dioxide and accompanying impurities and to the water in which the humic acids are dispersed. Roa (1928) compared the solvent effect of humic acids, carbonic acid, alkali carbonates, and sulfuric acid on basalt from the Giant's Causeway. Alkali carbonates, carbonic acid, and humic acid solutions at a controlled temperature between 32° and 52°C for 9 months decreased the silica and increased the alumina content above that in the original basalt. These reagents and probably also magnesium bicarbonate are effective in desilication, in which silica is removed from the parent rock, apparent porosity and open spaces are increased, and the alumina content in the resulting material is concentrated to form hydrous aluminum oxides.

No apparent relation exists between the presence or thickness of lignite and the depth and quality of the bauxite in Arkansas. Thick deposits of high-grade bauxite lack overlying lignite, and poor-quality bauxite may underlie the thickest lignite deposits. In Georgia and Alabama the bauxite deposits are not overlain by lignite, which is seldom present in the clay formations from which the bauxite was formed. In the underclays of Pennsylvanian age in Illinois, Missouri, Ohio, and Pennsylvania the best-quality underclay is usually the center of the clay and not the upper part. The diaspore clays of Missouri are related to fractures rather than to coal or lignite seams. The diaspore clays of Pennsylvania were derived from spotted clays near the center of the section rather than the clays near the top. These relations suggest that the presence of lignite or coal may have contributed carbonic acid to ground-water solutions that diffused through the upper impure clays and gathered alkalies and magnesium as the carbonates and bicarbonates. When these solutions reached the central zone, their concentration was sufficient to attack the clays in which original depositional structures provided a zone of permeability that permitted the saturated solutions to be removed and a fresh supply of reagents to enter. At many localities the carbonic acid that altered clay to hydrous aluminum oxides came from rain water or from the complete decay of plants not recorded as lignite or coals. Campbell (1910) emphasized the abundance of carbon dioxide in the spring waters of the tropics and its effectiveness in removing silica from silicate minerals in weak aqueous solutions of alkaline carbonates and in the concentration of alumina as bauxite. Veatch (1909) reasoned that the bauxite of Georgia was formed by carbonic acid waters of meteoric origin becoming alkaline and the alkaline carbonates removing silica as they circulated through the white clays of the Tuscaloosa formation.

Surface Weathering in a Warm Moist Climate

Many writers are convinced that laterites and bauxite are the products of surface weathering requiring special topographic and climatic conditions. Mead (1915) considered the bauxite of Arkansas to be formed by surface weathering

of syenite under a tropical or subtropical climate. The absence of frost allows the preservation of openings along which solutions remove silica and concentrate alumina as gibbsite with the preservation of the structure of the feldspars of the syenite.

Maclaren (1906), Simpson (1912), Holmes (1914), Woolnough (1918), Fox (1923), and others recognize the important result of alternating wet and dry seasons, like those that exist in the tropics, on the decomposition of rocks and the formation of bauxite and laterite. The infiltrating water should remain for long periods of time in contact with the rock. During the wet season, the interspaces of the upper part of the profile are kept filled, but during the dry season these are practically dry. The percolating waters remove certain substances in solution, while the evaporation of the water brings other substances to the surface by capillarity.

Harder (1949) concluded that bauxite deposits are significant stratigraphically because they mark long periods of emergence, quiescence, nondeposition, and subaerial weathering.

If the process of surface weathering is to be applied to the formation of bauxite, the deposits must be exposed at the surface for the agents of weathering to operate on them. This exposure would necessitate the recognition of disconformities or time breaks at the top of the bauxite deposits now covered by clay deposits. Shearer (1917, p. 124) states that over wide areas the upper surface of the bauxite of Georgia is an unconformity. Rettger (1925) recognized that the position of the bauxite of Alabama is controlled largely by the disconformity at the base of the Wilcox group and the old surface of the eroded Clayton formation. But many of the bauxite deposits of Georgia and Alabama show no evidence of a break between the bauxite and the overlying clay. Deposition of clays can take place on a surface that has been exposed to weathering without leaving any evidence of the time interval. This is especially true where the clay minerals deposited when sedimentation is resumed are the same as those in the underlying beds. An opportunity to test this possibility was presented in 1942 when several geologists accompanied the writer to the Gladding McBean pit, Lincoln, California. In 1928 the disconformity between the kaolinitic clay of the Ione formation of Eocene age and the clays of the rhyolitic beds of Oligocene and Miocene age was photographed (Allen, 1928, Pl. 26b). The face shown in this photograph was moved by mining the clay, and new faces were exposed in 1942. The geologists were told that two series were exposed in the pit, the upper one of Oligocene or Miocene age containing the reworked minerals of the lower series plus some montmorillonite minerals derived from the alteration of volcanic ash. They were asked to indicate where the disconformable contact between the two series should be drawn. Where massive clays of the upper series rested on clays of the Ione formation lacking well-developed bedding, no agreement was reached, and no one was convinced of any erosional break. But in another part of the pit, where the upper series filled erosion channels that cut prominently bedded clays of the Ione formation, they agreed that here was evidence of an erosional break separating the two series. It must be admitted that time breaks between bauxite or diaspore deposits and the overlying clay can exist without the proof needed for their recognition. The writer does not imply that he assumes a disconformity is present at the top of every bauxite and diaspore deposit, but he is merely stating the possibility that some disconformities may exist which he and other geologists have overlooked.

McQueen (1943) has described high-grade fire clays in Missouri that were formed on weathering surfaces, and overlain by low-grade clay, with the contact between the two obscure and in some places brought out only as a result of detailed stratigraphic work in the area.

Several investigators are convinced that plagioclase and nepheline weather directly to bauxite in tropical regions. Aluminous laterite and bauxite are reported to form directly by the weathering of igneous rocks above the water table under favorable conditions, and clay minerals are formed below the water table either by direct crystallization or by resilication of aluminum hydroxide (Goldich, 1948). Thin sections of a basic igneous rock from the bauxite district of British Guiana from the collection of C. S. Ross show a transition

zone from the feldspar to gibbsite in a fraction of an inch. Evidence of the formation of bauxite directly from feldspars and nepheline has not been observed at the localities investigated by the writer, and it does not seem to be the dominant process at these deposits. Furthermore, certain difficulties that prevent the general acceptance of this process are mentioned here. In the series of six specimens from a boulder through a zone 1½ feet thick to the surrounding bauxite of commercial grade in the Neilson under-ground mine of Arkansas, more kaolinite is present in the feldspar laths at its core (Pl. 1, fig. 6) than at its margin. If the feldspars were altered directly to gibbsite, the silica to form this kaolinite would have to pass through the gibbsite at the margin to reach the core and change the gibbsite to kaolinite. Why then was not the gibbsite at the margin also resilicated to kaolinite, and why was partial resilication restricted to the core? At the high-iron bauxite or laterite deposit of Oregon (Fig. 2), is one to believe that from the surface to a depth of 170 feet all the silica was removed from plagioclase to form gibbsite, and later silica from an unknown source traveled downward through 60 feet of gibbsite without resilicating it but resilicating 100 feet of the underlying gibbsite to form the transitional clay layer? This is not in keeping with the law of conservation of energy in nature nor with the petrographic evidence at the deposit.

On the other hand, the known conditions do not exclude the possibility that clay minerals can form below the ground-water table by direct crystallization at the same time as bauxite forms near the surface where drainage and other conditions are favorable. Also, it seems possible for streams to cut downward and headward in a region where bauxite is forming at the surface and lower the ground water table. Then bauxite could form from clay minerals as good drainage conditions were established at increasing depths. Progressive removal of silica not only offers a satisfactory explanation for the formation of bauxite but also for the clay present in a bauxite deposit. Moreover, petrographic evidence indicates that desilication to the clay-forming stage was the usual process at the deposits examined and that at only a few was it carried to the stage where aluminum hydrates were formed.

Ground Water

The writer has observed water seeping from a clay bed in the Columbia River basalt exposed along a valley. The original structures of some clays are sufficiently permeable to permit diffusion of ground water through them. If these waters contain carbon dioxide or alkaline carbonates, they would remove silica from clay minerals, as has been demonstrated by experiments (Allen, 1935a). Spring waters flowing from the surface of clays of Georgia are said to be opalescent and high in silica. Campbell (1910) reports that spring water in the tropics is more highly carbonated than in higher latitudes and that water issuing from underground at all seasons gives off carbon dioxide upon exposure to the air when ferric hydrate is deposited. These observations imply that surface exposure of the clay bed is not necessary during alteration. As long as an inlet for an adequate supply of carbon dioxide water and an outlet for the removal of the dissolved and dispersed substances are provided, the resulting concentration of the alumina and increase of porosity would form bauxite or diaspore clay. The sandstone-lined sink holes of Missouri that provided an outlet for ground-water solutions allowed silica to be removed from flint fire clay along fractures and the permeable parts of the clay and concentrated alumina to form diaspore clay (Allen, 1935a).

Boehmite

In the laboratory, boehmite has been prepared under a variety of conditions.

Heating experiments by Laubengayer and Weisz (1943) indicate that gibbsite and bayerite are stable below 155°C, that boehmite is stable between 155°C and 280°C, and that diaspore is stable between 280°C and 450°C.

Weiser (1935) reports that the gelatinous precipitate obtained by neutralizing an aluminum salt solution with ammonia is boehmite. Aging of the precipitate in water containing alkali results in a growth of crystals that gradually change boehmite → bayerite → gibbsite which requires an increase in water content.

Boehmite has been prepared by partially dehydrating gibbsite by heating it under pressure. Laubengayer and Weisz (1943) made the boehmite from gibbsite by holding it in a bomb under hydrothermal conditions at 350°C for several days. Schwiersch (1933) found that gibbsite when heated loses water to form boehmite, which continues to lose water to form gamma-Al_2O_3. Lehl (1936) states that boehmite is formed by slow thermal disintegration of gamma trihydrate of aluminum oxide and that well-crystallized boehmite is very stable and shows no noticeable signs of aging.

Good crystals of boehmite have been formed by heating $(AlNO_3)_3$ solutions made acid with HNO_3 at 320°C to 360°C and 200 to 300 atmospheres of pressure (Weiser, 1935, p. 94).

De Lapparent (1936a) believed that the formation of natural boehmite is favored by long contact with humic acid solutions and that diaspore is formed during subsidence when the overlying load of sediments causes a rise in the isogeotherms. Boehmite is associated with carbonaceous sediments at some localities such as in the shales of the Puget group of Washington and locally in Missouri, but so is diaspore in the clays of Pennsylvania and Missouri. The presence of boehmite in the same specimen with diaspore or gibbsite indicates a lag in the response of these minerals to changing conditions and the persistence of the metastable phase after the conditions have changed. Gruner's experiments (1944) are of interest in this connection. He formed boehmite in acid solutions where Al ions are present in excess of those required for pyrophyllite at temperatures up to 400°C, considerably above the 280°C given by Laubengayer and Weisz as the upper stable limit of boehmite. Noll (1935) found that mixtures of Al_2O_3:SiO_2 = 1:2 yielded kaolinite when heated to 250°C to 400°C at a pressure of 300 atmospheres, but mixtures richer in alumina heated under the same conditions formed boehmite in addition to kaolinite.

The laboratory conditions under which gibbsite, boehmite, and diaspore were synthesized are difficult to apply to field relations. Diaspore, the highest-temperature mineral of the group, has been reported to form as a result of metamorphism (De Lapparent, 1933). Laubengayer and Weisz (1943) report that diaspore is stable between 280°C and 450°C, but the only diaspore formed in the laboratory consists of secondary growths around seeds of diaspore that were placed in a bomb. In the absence of diaspore seeds, boehmite forms at temperatures up to 400°C. The minerals present in the flint clays of Missouri and their lack of deformation offer no proof of high temperatures. Furthermore, boehmite and diaspore occur in the same hand specimen and were subjected to the same temperatures during alteration. De Lapparent's suggestion (1936a) that the formation of boehmite is favored by long contact with humic acid solutions does not offer an explanation, because diaspore and boehmite occur in the same pit within a few inches of each other and within inches of the same coal seam. From present relations, formation of diaspore also appears to be favored by humic acid solutions, but from these observations one cannot reconstruct the conditions in the chemical system at the time when the diaspore and boehmite were forming.

Somewhat similar difficulties apply to the gibbsite and boehmite which occur in the same hand specimens collected from the deposits in Arkansas and California. The formation of gibbsite in the laboratory is favored by alkaline solutions and by the rate at which carbon dioxide is introduced into an alkali-aluminate solution. If the carbon dioxide is allowed to act slowly at ordinary temperature and the container is permitted to stand exposed to air, gibbsite is obtained. If the carbon dioxide is introduced rapidly, amorphous aluminum hydroxide is formed. Temperature modifies the conditions of crystallization, because gibbsite of coarser crystalline structure is obtained by allowing carbon dioxide to pass slowly over a solution of aluminate heated to 96–100°C (Lehl, 1936, p. 47). Precipitates of boehmite, according to Weiser (1935), age in cold water containing alkali and change to gibbsite, which is more stable than boehmite or bayerite. Well-crystallized boehmite is very stable and shows no signs of aging (Lehl, 1936). The formation of boehmite is favored by acid solutions, by temperatures of 250° to 400°C, by a pressure of 300 atmospheres, and by an excess of Al ions over those required to form hydrous aluminum silicates (Gruner, 1944; Noll, 1935).

Boehmite can also be formed from gibbsite by heating it under pressure at 350°C (Laubengayer and Weisz, 1943). When coarsely crystalline, the metastable phase can persist under conditions where fine-grained material is transformed to the more stable form.

Conclusions

Desilication is the most important process in the formation of bauxite, diaspore, and boehmite deposits. The removal of silica from feldspars, from igneous rocks, and from clays increases the alumina and titania as well as the porosity of the resulting rock and facilitates the retention of the structures of the original materials. The presence of undistorted shapes of feldspar crystals in rocks now composed partly or wholly of clay minerals suggests that the clay minerals were formed by desilication, because the volume increase which accompanies the substantial addition of new substances tends to distort the original shapes. In many bauxite and diaspore deposits, clay minerals are formed early in the process of desilication and are intermediate products in the formation of bauxite and diaspore.

At the localities studied, gibbsite, diaspore, and boehmite are not characteristic of or limited to any particular geologic period or to any particular type of parent rock, but are related to periods of weathering, some of long duration and great lateral extent. The gibbsite of Georgia, Alabama, Virginia, and Arkansas was probably formed during the Eocene epoch. Gibbsite was formed in California during the Eocene; at Castle Rock, Washington, during the Eocene; at Molalla, Oregon, during the Miocene; and in the high-iron bauxite district of Oregon during the late Miocene or early Pliocene.

The diaspore deposits of Missouri and Pennsylvania are of Pennsylvanian age; and one occurrence of diaspore in the Cle Elum iron deposit, Kittitas County, Washington, is of Eocene age.

The boehmite of Missouri and Kentucky occurs associated with flint clays of Pennsylvanian age, and the boehmite of Arkansas, Washington, and California was probably formed during the Eocene epoch.

The parent rock of the diaspore of Missouri, Pennsylvania, and Washington is a flint clay composed of finely divided kaolinite having low plasticity and breaking with conchoidal fracture. The gibbsite and bauxite of Georgia and Alabama was formed from a kaolinitic hard to semihard clay. The chief difference between the two kinds of clay is that the flint clay is made up of very fine particles that lack the degree of symmetry present in the stacking of the molecular sheets of kaolinite. Whether this would cause sufficient differences in the chemical system to bring about the crystallization of diaspore rather than gibbsite is not known. Boehmite has formed from flint clays in Missouri, Kentucky, Washington, and California. The occurrence of the monohydrates of alumina, diaspore and boehmite, as alteration products of flint clay seems to be the rule in the districts studied during this investigation.

The factors controlling the formation of gibbsite, boehmite, and diaspore in nature probably include: (1) the chemical system, (2) reagents available and whether alkaline or acid, (3) reaction rate or time, (4) temperature, (5) pressure, (6) drainage.

Bauxite, boehmite, and diaspore clays are formed under nearly similar climatic conditions, in which a warm moist climate and thorough leaching under conditions of good drainage allow the removal of silica from aluminum silicates and clays. Chemical decomposition is relatively rapid and proceeds to a stage where hydrous aluminum oxides are formed in tropical and subtropical regions. Decomposition is partly caused by the slightly higher temperatures which increase the rate of chemical reaction. Robinson (1932) believes that a rise of 10°C corresponds approximately to doubling the velocity of the chemical reaction. Also, the abundance of vegetation and carbon dioxide that is brought down during heavy rains saturates ground water with carbonates and bicarbonates that cause desilication of the parent rocks. Removal of silica increases porosity and permeability and accelerates the process as the number of open spaces and their connections multiply. The available evidence indicates that the climate of each period when gibbsite, boehmite, and diaspore were formed

in the United States was warm and moist. At many localities decomposition took place at the surface, but later this surface was covered by clay which either marked its position by a disconformity between it and the overlying kaolin deposits, or concealed any visible sign of the erosional break. At some localities, ground-water solutions containing ample carbonic acid and alkaline carbonates moved along relatively permeable zones in the clays and effected desilication under conditions of good drainage.

The second process, the migration of clay minerals and hydrous aluminum oxides, not only fills the available openings and decreases permeability but also locally distorts and disturbs the original rock materials. The volume increase in some places is sufficient to destroy original and relict textures. The zone of compact clay between the bauxite and the syenite of Arkansas (Goldman and Tracey, 1946, p. 573) was probably formed by this process. Likewise, the clay near the core of the boulder-like masses (Goldman, 1949, p. 1890) which lacks relict texture of the feldspars and shows distortion may represent late clay that migrated inward through the porous gibbsite zone at the edge. The deposition of the late clay was localized by an early impermeable clay zone that prohibited further inward invasion. The clay near the core of the boulders which retains the shapes of the feldspars was probably formed by desilication during the early stage of bauxitization. The addition of late clay by migration masks the transitional zone at some places and confuses the relationships. The addition of hydrous aluminum oxides to cracks and open spaces increases the alumina content of the resulting rock. The tubes, nodules, and irregularly shaped masses formed in this way may lack any evidence of the original structure of the parent rock. This process is aided by conditions of good drainage, but its effective operation must await the formation of openings by desilication.

The third process, resilication, operates under conditions of poor drainage in low-lying areas subject to fluctuations of ground-water level. It causes less distortion of rock textures than the migration of clay minerals because the process involves the substitution of silica for some water and a liberation of the surplus alumina of gibbsite above that required to form kaolinite. The structure of the original rock and the apparent porosity have been preserved in some of the resilicated specimens examined during this investigation.

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Manuscript Received by the Secretary of the Society, September 25, 1950.

Published by Permission of the Director, U. S. Geological Survey.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 63, PP. 689-700, 1 FIG., 4 PLS. JULY 1962

PIMPLED PLAINS OF EASTERN OKLAHOMA

BY MAXWELL M. KNECHTEL

ABSTRACT

Patterns formed by the networks of furrows separating the natural mounds of pimpled plains in eastern Oklahoma are regarded as attributable to shrinkage-polygon systems of coarse texture comparable to those occurring (1) in mound-studded parts of the northwestern United States that are underlain by vertically jointed basalt, (2) in tracts of Alaskan tundra that are occupied by ice-wedge networks, and (3) in beds of playa lakes in the arid southwest that are cut by systems of giant desiccation fissures. While the intermound furrow networks characteristic of pimpled plains in the mid-continent region may owe their origin to former frigid climatic conditions, one of a number of alternative possibilities is that the peculiar configuration of these surfaces, as exemplified in eastern Oklahoma, has resulted from erosion of systems of giant soil polygons caused by desiccation. The time of origin of the pimpled plains of eastern Oklahoma is believed to have been not earlier than late Pleistocene.

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INTRODUCTION

Low-relief topography, or microrelief, formed of innumerable closely spaced small hummocks or mounds (Pl. 1, figs. 1, 2), is a conspicuous feature of many large tracts of nearly level to gently sloping land in some states west of the Mississippi River, as well as in some other parts of the world. Figure 1, based on a map constructed by Price (1949, Fig. 1), shows a number of physiographic provinces wherein such mounds are reported to occur in the United States.

Such mounds are known as "prairie mounds", "hog-wallows" (California), "Mima mounds" (Washington), and "puffs" (Australia). The origin of the mound-studded surfaces, which in this country are commonly called "pimpled plains" (Fenneman, 1931) has been a subject of much discussion for more than 100 years and has been attributed to various natural processes. Interpretations of their genesis set forth in the references cited involve such concepts as erosion by networks of rivulets (Le Conte, 1874, p. 366; Waters and Flagler, 1929; Melton, 1929; Knechtel, 1949); erosion by wind (Ellis and Lee, 1919, p. 29–30; Rich, 1934); "mud-volcano" phenomena (Hobbs, 1907); hydrostatic pressure of ground water (Nikiforoff, 1941); "concretionary action" (Branner, 1905); frozen-ground phenomena (Bretz, 1913, p. 105; Newcomb, 1940; Péwé, 1948); construc-

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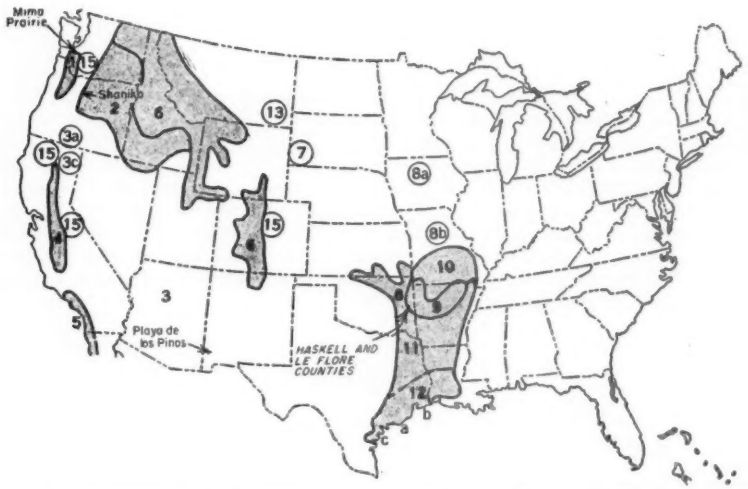

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> "Many of the mounds occur within areas of bedrock exposure but their areal distribution bears no direct relation to that of any of the different bedrock units. Some of them occur also on ancient gravel terraces and others on the higher parts of the Recent alluvial plains along the larger streams. . . . Those in Le Flore County occur at altitudes ranging through several hundred feet though they are present only on nearly level surfaces and gentle slopes. For example, on the gently sloping upper surface of a prominent hogback a mile south of Bokoshe mounds occur approximately 400 feet above the Arkansas River level; within half a mile of these and 300 feet lower are others on the higher parts of the alluvial plain along Buck Creek.
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Although some of the elementary polygons of the ice-wedge and desiccation-fissure systems pictured are much larger than the subcircular or polygonal mound-relief patches in eastern Oklahoma (Pl. 2, fig. 1), the patterns are much alike. Notable points of resemblance which seem to indicate kinship among all these patterns are (1) a tendency of the polygonal and subcircular elements of all such patterns to line up in rows; and (2) imperfect latticelike cross-alignment of many of the rows. Fracture systems due to contraction commonly show similar patterns; inasmuch as such patterns—where related to ice wedges, desiccation fissures, or columnar basalt—do appertain to fractures due to contraction, it is considered that this may be true also of the eastern Oklahoma intermound furrow patterns. Contractive forces that can produce patterns like those illustrated are apt to result from cooling or desiccation, and these agencies play the leading role in three variants of the fracture-polygon hypothesis involving (1) fractures in bedrock, (2) fractures in frozen ground, and (3) fractures due to desiccation. The question is: Which of these three categories of fracture phenomena could have been responsible for the eastern Oklahoma furrow patterns?

Jointing in Bedrock

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In area centering approximately 5 miles north of Stigler, Haskell County, Oklahoma. (Photo by Production and Marketing Administration, U. S. Department of Agriculture.)

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Mima prairie, 1½ miles west of Little Rock, Thurston County, Washington. (Photo by U. S. Geological Survey.)

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FIGURE 3

FIGURE 4

FIGURE 1

FIGURE 2

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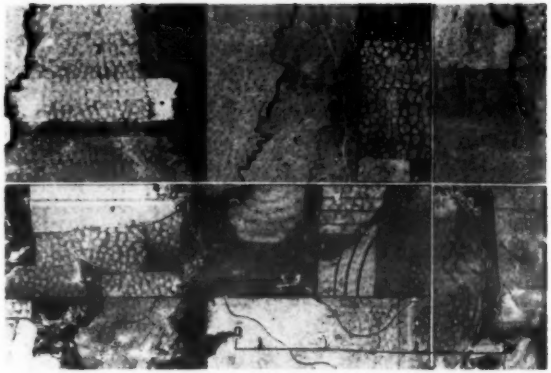

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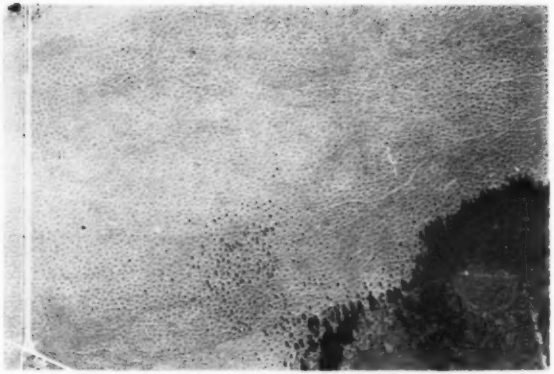

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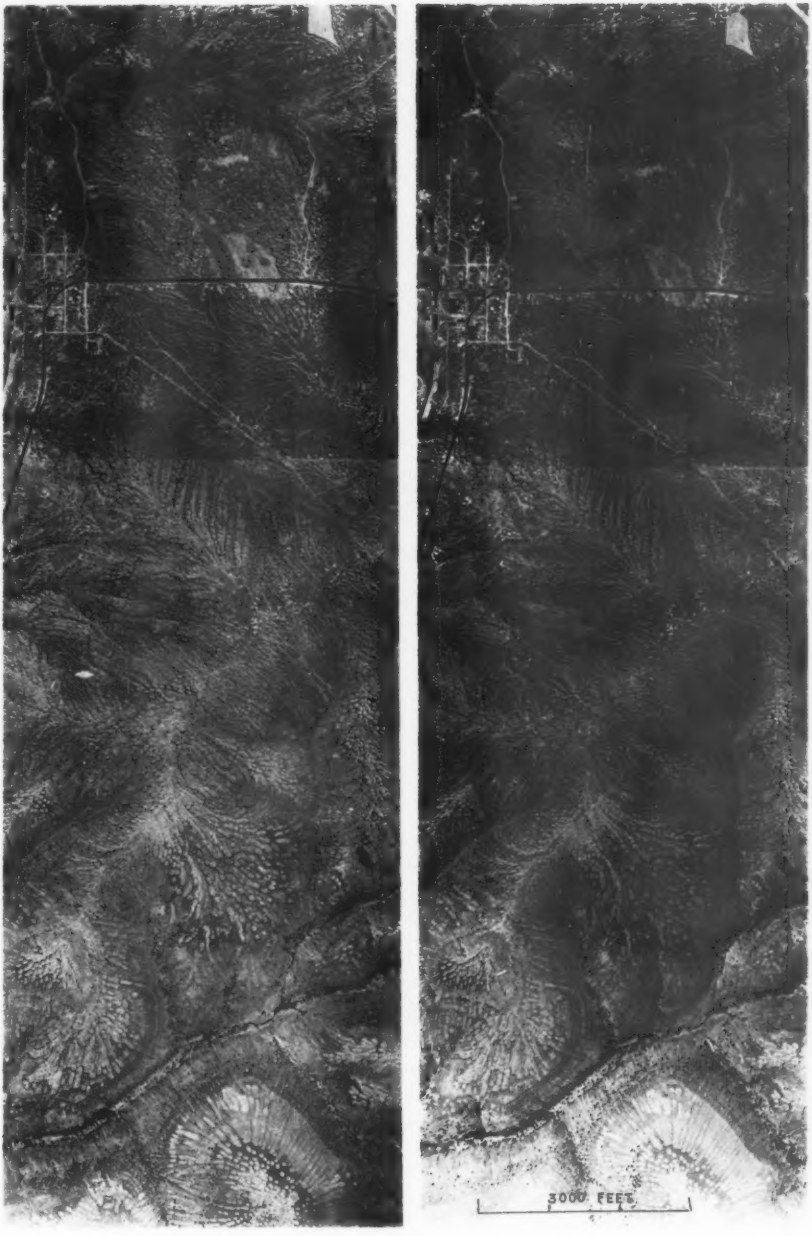

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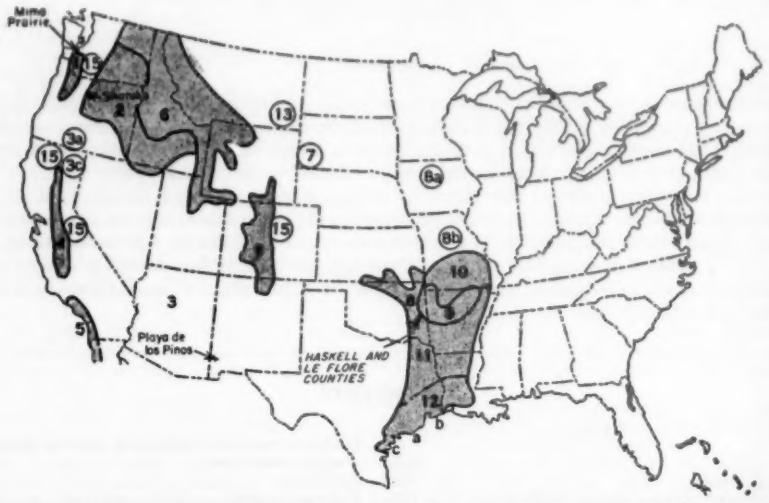

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FIGURE 3

FIGURE 4

FIGURE 1

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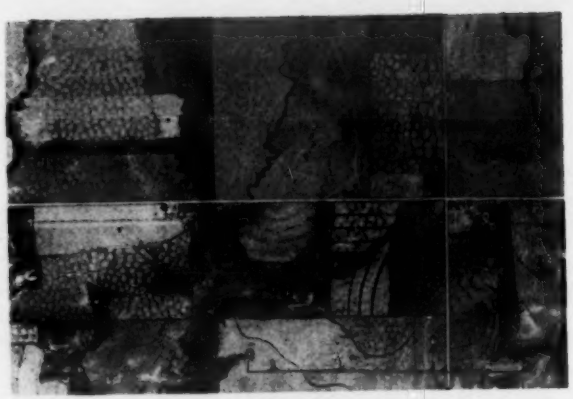

FIGURE 1

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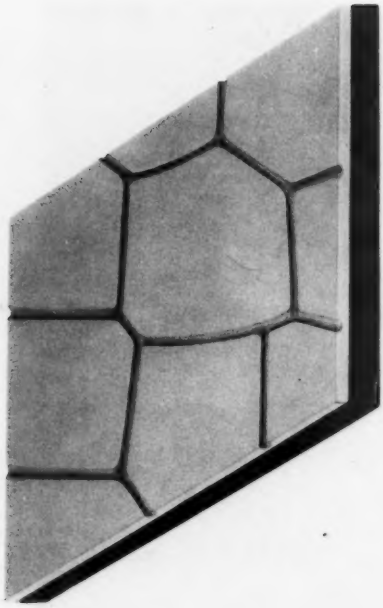
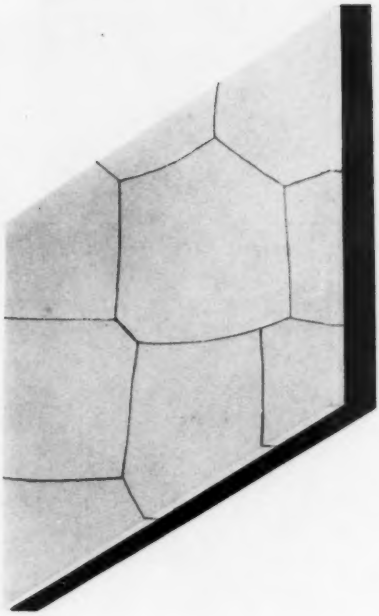


FIGURE 3

FIGURE 4



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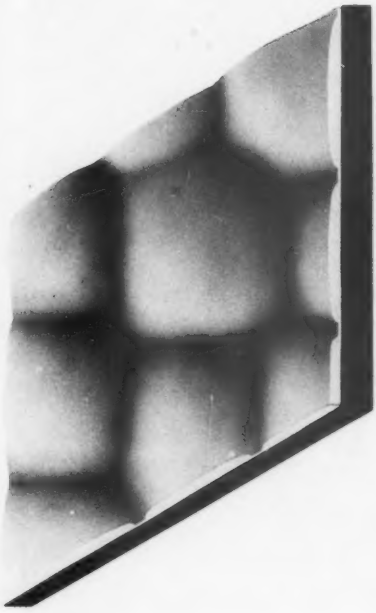

FIGURE 2

POSTULATED STAGES IN EVOLUTION OF EASTERN OKLAHOMA MOUND RELIEF

Drawings by O. W. Goodhue

With reference to intermound furrow patterns at a number of localities on the Columbia River plateaus, in Washington and Oregon, where mound relief has developed in blankets of volcanic ash resting on basalt, Waters and Flagler (1929, p. 223) have commented as follows:

> "The intervening areas between the mounds form a definitely integrated and minutely adjusted drainage system. The symmetrical development of this drainage system may be influenced by the columnar jointing of the ash. The importance of this factor is problematical, however."

Their photograph of a locality along the margin of Deschutes Canyon near Maupin, Wasco County, Oregon (Fig. 3 of their paper), shows narrow linear furrows dividing the thin ash cover into strips and, as they point out, "some of the strips have been segmented into mounds". Their picture suggests that the straight, rather uniformly spaced furrows between the strips lie above widely spaced vertical fissures in basalt that is visible under the ash. Joint systems with a coarse texture like that of these intermound furrow patterns appear on several stereophotos of other areas in the same region (Frost and Woods, 1948, Figs. 69–73) that are underlain by basalt. These stereophotos and others (Pl. 3) showing the neighborhood around Shaniko, Wasco County, tend to confirm the impression given by the Deschutes Canyon picture and seem also to establish that the furrows between segments of the strips are likewise an expression of cracks in the basalt floor. In the walls of the canyon that cuts across the eastern end of the area depicted in the Shaniko stereopair, systems of parallel, evenly spaced joint planes trend in several directions. The basalt of that locality is apparently divided by intersecting joints into columns many feet in diameter, the tops of which are inferred to have, in aggregate, the pattern of a great system of polygons. In the gently sloping ground just above the rim of the canyon are many evenly spaced, straight furrows which appear to follow the systems of parallel joints trending most nearly at right angles to the topographic contours, that is, in the direction of slope. Accordingly, the furrows on the prominent, rounded spur east of the canyon show a distinct radial pattern, as do also those of mound-studded areas in other parts of the area. The intermound furrow patterns in the vicinity of Shaniko seem therefore to be genetically associated with columnar jointing in the underlying basalt.

A similar genetic association of natural mound systems with fractures in bedrock seems possible for mounds in the Arbuckle Mountains of south-central Oklahoma. W. E. Ham (Oral communication, 1951) has informed the writer that these mounds rest on a floor of vertically jointed cherty strata of the Arbuckle limestone (Cambrian and Ordovician). The mounds are composed of residual chert and sandy soil, and are restricted in their occurrence to those parts of the Arbuckle limestone in which chert is abundantly present in the bedrock.

Horberg (1951, p. 6) mentions, but doubts, the possibility that the intersecting ridge patterns in the bed of glacial Lake Agassiz, in North Dakota, which he regards as fracture patterns, may have been caused by "tectonic adjustments", involving the underlying bedrock, that accompanied isostatic upwarping upon retreat of the continental glacier.

In view of the apparent relation between patterns characteristic of some intermound furrow systems and fracture systems in the underlying bedrock, the possibility of a similar relation for the pimpled plains of the mid-

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(Drawings by O. W. Goodloe).

FIGURE 1.—BLANKET OF SOIL RESTS ON LEVEL OR GENTLY SLOPING FLOOR OF FIRMER MATERIAL
Cracking due to shrinkage has divided soil into large polygonal blocks. (Area represented is approximately one acre.)

FIGURE 2.—SUBSIDENCE ALONG FISSURES BETWEEN BLOCKS HAS PRODUCED NARROW GROOVES ANASTOMOSING AMONG LOW MESALIKE UNITS

FIGURE 3.—RIVULETS FORMED IN SUBSIDENCE GROOVES MODIFYING FORM OF MESALIKE BLOCKS

FIGURE 4.—CORNERS OF BLOCKS WORN AWAY BY CONTINUED EROSION
Leaving their centers left standing as low, subcircular mounds.

continent region should not, perhaps, be entirely discounted. With reference to patterns visible on an airphoto of mound relief near Tyler, Smith County, Texas, Melton (1929, p. 126–127, Pl. IV) has called attention to:

> "numerous gullies which are advancing up the slopes from the larger streams (and are) dissecting the light-colored surface soil into long strips with a certain degree of parallelism. These long strips of light surface soil are obviously being further dissected and cut into numerous smaller segments by a series of gullies of the second order. . . . Several other pictures which show this same relationship between rows of mounds and long straight gullies have been found".

Such a relationship is reminiscent of that apparently existing between the furrows, strips, and mounds near Shaniko, Oregon, where the patterns seem to be attributable to columnar jointing in the Columbia basalt.

However, the soil materials exhibiting the patterns in eastern Oklahoma are spread over outcrops of folded and faulted Paleozoic sedimentary rock units ranging in dip from very steep to horizontal and differing greatly, *inter se*, in their lithologic make-up and hence in their probable behavior under deformational stresses; also, the soil materials rest at many places on undisturbed late Tertiary and Quaternary alluvial deposits. It is scarcely conceivable that furrow systems so uniform in texture as those of the pimpled plains would result from fracturing of bedrock that differs so greatly, from place to place, in lithologic characteristics and in geologic structure. Moreover, intersecting vertical or subvertical joint systems comparable to those of the Columbia basalt or the Arbuckle limestone, if present, would probably be a conspicuous feature of the bedrock outcrops where the pimpled plains occur; comparable bedrock fracture systems are not known to the writer to be associated with these surfaces anywhere in eastern Oklahoma.

Fissures in Frozen Ground

According to Newcomb (1940) and Péwé (1948) the furrow patterns of the Mima mound systems near Puget Sound, Washington (Pl. 2, fig. 3), show evidence of inheritance from ice-wedge networks comparable to those illustrated (Pl. 2, fig. 4. *See also* Taber, 1943, Pl. 21, fig. 1; Washburn, 1947, Pl. 28, fig. 1; Leffingwell, 1919). Such a possibility has to be considered in dealing with the origin of mound relief in places where sufficiently low temperatures may be assumed to have existed at the time of origin, as in the periglacial zones marginal to the continental ice sheets of Pleistocene time. Lozinski (1933, p. 21) has even suggested—without reference, however, to ice wedges—that mound-relief systems as far south as the Gulf Coastal Plain of the United States may be frozen-ground phenomena comparable to mound fields, known as *Palsen*, occurring in the Polish Carpathian foreland. According to Sharp (1942) the *Palsen* of tundra soils in northern Europe and Siberia are probably "products of differential freezing and thawing that has caused upwelling of fine material". However, the intersecting minor ridges in the Lake Agassiz basin, North Dakota, which Horberg (1951) has tentatively interpreted as frozen-ground phenomena, are regarded by Nikiforoff (1952) as more probably related in their development to wave action or other causes. Agencies that are not dependent on any limited range of temperature seem to call for special consideration with reference to the origin of pimpled plains in areas far from any present or past glaciation, particularly where the present climate is temperate to warm.

Fissures Due to Desiccation

The contraction that produces ice-filled fissures in tundra is believed (Leffingwell, 1919; Black, 1951) to be due to cooling at extremely low temperatures, whereas that which commonly produces vertical joints in basalt results from cooling at very high temperatures. Giant playa fissures, which may form at various temperatures, are due primarily to drying. The pimpled plains of eastern Oklahoma are not associated with basalt or other igneous rocks and, consequently, if we admit (1) that the intermound furrow systems of those plains are in reality a result of contraction, (2) that no such extremely cold climatic conditions as those of the Arctic tundra environment existed in eastern Oklahoma when the systems were initiated, and (3) that chemical reactions that might cause shrinkage are not involved, it follows that the contraction was there most

probably due to desiccation that may have taken place under climatic conditions different from those of today.

Observations tending to support the suggestion that these furrow patterns represent extensive systems of desiccation-fissure polygons are on record for localities in various parts of the world. The great desiccation-fissure systems on Playa de los Pinos, New Mexico, already mentioned (Pl. 1, fig. 4; Pl. 2, fig. 2), are comparable in pattern and coarseness of texture not only to some of the ice-wedge networks in northern Alaska (Knechtel, 1951) but also to the intermound furrow systems of the Oklahoma pimpled plains. Regarding patterns somewhat similar to those of the Oklahoma mound fields, appearing on an airphoto of a locality along Red River in Cass County, southeastern North Dakota, within the southern extremity of Glacial Lake Agassiz, Frost and Woods (1948, p. 19, fig. 6) have commented:

> "The left half of this area exhibits the pattern characteristic of organic lacustrine clays. The organic topsoil is approximately 20 inches deep, and when dry, contains long deep shrinkage cracks."

Ellis and Shafer (1928, p. 241–248) have described shrinkage polygons farther north in Lake Agassiz, which they attribute to desiccation. According to Horberg (1951, p. 14), residents of Pembina County, North Dakota, report "desiccation cracks up to 6 inches in diameter and over 10 feet in depth" in the clays of Lake Agassiz. He observes that the occurrence in the same vicinity of soil wedges in clays and silts supports the view that these wedges have formed in desiccation cracks, though he doubts this possibility on account of the large size of the wedges and their upward-flaring cross section. He doubts also (1951, p. 13) that desiccation was responsible for openings in the sediments of Lake Agassiz that are occupied, near Concrete, North Dakota, by gravel wedges which he regards as probably related in origin to ground ice. He asserts that desiccation cracks "would not be so large or have flaring walls, nor would they develop in permeable gravels". The size of both the soil wedges and the gravel wedges nevertheless appears to be consistent with the possibility of their origin through desiccation, inasmuch as the desiccation cracks in the bed of Playa de los Pinos, New Mexico, are comparable in width and are even more widely spaced. Moreover, fillings that flare upward are apt to result from caving of the walls of desiccation cracks that open repeatedly and, notwithstanding the permeability of the beds that are cut by gravel wedges, these beds, consisting largely of "flat pebbles of Cretaceous shale with average maximum diameter of about 2 inches, sandy matrix, silt and sand layers" (Horberg, 1951, Table 5, p. 12), evidently contain considerable argillaceous matter which, if sufficiently colloidal in its behavior toward water, would favor cracking through desiccation.

The microrelief of the crab-holey country of Australia, which is characterized by alternating "rises", or *puffs*, and depressions known locally by the names *crab hole*, *melon hole*, *gilgai*, *Bay of Biscay*, and *devil crab*, is related (Prescott, 1931, Pl. III, fig. 1, p. 15) to "alternation of wet seasons or floodings on the one hand and of droughts on the other". Prescott says further (1931, p. 55) that the crab-holey ground of an area in New South Wales consists of heavy alkaline alluvial soil "which readily cracks in dry weather".

The occurrence in the Gobi desert of Mongolia of "certain desert features that more or less resemble periglacial ones" is described by Hörner (1950), who asserts that "certain desert cracks might be misinterpreted as (cracks formerly occupied by) ice wedges". He emphasizes the need for careful study of the likenesses and also of possible distinctions between certain periglacial and certain arid-region phenomena.

In meadows near the village of Zakrzew, Poland, "a regular polygonal (penta- and hexagonal) network of crackings", 1.5 to 3 meters in diameter is reported (Jahn, 1950, p. 156) as having resulted from rapid drying of a silt layer that had formed under water. The "crackings" are described as "very similar to the polygonal network found in the Arctic tundra (e.g., to the fissure polygons in Greenland and Spitzbergen). ... This phenomenon proves that the markings formed through drying are similar to those which are produced by the action of frost".

Possible Causes of Development of Mounds

General Statement

Assuming that great systems of fracture polygons provided the framework for development of the pimpled plains in eastern Oklahoma, the microrelief that characterizes such surfaces may have evolved in several stages (Pl. 4). Of interest with reference to the sequence of events postulated are three previously published suggestions, relating to areas outside eastern Oklahoma, in which expansion, subsidence, and erosion are advanced, respectively, as contributory causes of surface configuration comparable to that of the pimpled plains. All three of these possibilities enter into the present analysis, though only subsidence and erosion are invoked in the writer's hypothesis (Pl. 4).

Expansion

Surfaces characterized by irregular hummocks 3–5 or more feet in diameter occur on the plain adjacent to the Red River in Manitoba (Ellis and Shafer, 1928). The hummocks consist of soil derived from "alluvium and shallow lake deposit laid over the deep laminated clays of Lake Agassiz". The depressions between the hummocks are underlain by iciclelike downward "intrusions" of dark soil that are described as having formed in systems of cracks caused by "enormous shrinkage on drying of soils high in colloids". The authors explain the hummocks as follows:

> "When the soil becomes wet, the wedged-in dark intruded material swells more than the adjacent material. Due to the plastic nature of the marly very fine sandy clay when wet, it reacts to the greater pressure developed by the intruded dark surface clay and heaving upward takes place where the surface clay is most thin."

While very small hummocks may be attributable to heaving due to swelling of clay material in systems of cracks, it is questionable whether this mechanism could by itself produce features as large as the pimple mounds of Oklahoma.

Subsidence

Péwé (1948) has suggested that the furrows between the Mima mounds near Puget Sound (Pl. 2, fig. 3) originated in essentially the same way as those separating mounds near Fairbanks, Alaska. He regarded the Fairbanks furrows as due to sinking of the ground upon melting of polygonal networks of ice wedges, leaving the centers of the polygonal blocks standing in relief. However, subsidence along the desiccation fissures on Playa de los Pinos, New Mexico (Pl. 1, fig. 4), has produced not mounds, but low mesalike blocks, the tops of which are horizontal planes that form obtuse dihedral angles with the walls of the interblock grooves. The rounded forms of the pimple mounds of Oklahoma probably have not resulted from subsidence alone, though subsidence may have played a part in their origin (Pl. 4).

Erosion

While intermound furrows inherited from tundra-fissure networks may have developed through subsidence of the ground upon melting of ice masses that had formed in the fissures, some other process seems to have been primarily responsible for the growth of such furrows in temperate to warm climates, where ground ice presumably could not have played so significant a role. Le Conte (1874, p. 366) pointed out that the material composing each unit mound is, apparently without exception, softer than the underlying material, though Péwé (Personal communication) states that the "mounds in the Fairbanks area, Alaska, have no hard underlayer". Le Conte attributed certain occurrences of mound relief in Oregon and California to

> "surface erosion acting under peculiar conditions. These conditions are a *treeless country* and a *drift soil* consisting of two layers, a fine and more movable one above and a coarser and less movable one below. Surface erosion cuts through the finer superficial layer into the pebble layer beneath, leaving, however, portions of the superficial layer as mounds. The *size* of the mounds depends upon the *thickness* of the superficial layer; the *shape* of the mounds depends upon the slope of the surface."

The process suggested by Le Conte was regarded by Waters and Flagler (1929) as having produced the mound relief of the Columbia River plateaus and has been invoked to explain mound relief in Texas by Melton (1929) and in eastern Oklahoma by Knechtel (1949).

Synthesis Involving Desiccation Fissures, Subsidence, and Erosion

The pimpled plains of eastern Oklahoma may have been created essentially by Le Conte's soil-erosion process, whereby similar surfaces are supposed to have been carved by erosion from a thin blanket of soil resting upon a floor of firmer material. The efficacy of that hypothesis seems nevertheless to be greatly enhanced by coupling it with the postulate that the "more movable" layer resting on the "less movable" material had become divided, prior to erosion, into large prismatic blocks.

Assuming that development of the intermound furrow systems began with formation of giant shrinkage-fissure networks caused by desiccation, the present configuration may have come about in essentially the manner illustrated in Plate 4. Though the fissures are therein represented as having developed only in the surficial blanket of soil, the intermound furrows may follow shrinkage cracks in the underlying claypan, or other firm material upon which this blanket rests, just as the intermound furrows of the Columbia River plateaus seem to follow systems of joints in the basalt floor beneath the blanket of ash from which the mounds of that region have been carved.

Mound development, as postulated (Pl. 4, fig. 2), begins with a system of grooves, similar to the grooves illustrated (Pl. 1, fig. 4), that have resulted from subsidence of the ground along desiccation fissures. As further postulated (Pl. 4, fig. 3), the furrows are deepened by rivulets that have formed along the fissures; because of the low gradient and small size of these rivulets, their transporting power is never great enough to clear away all the soil that is washed into them and the load of this material serves as a protective cover over the resistant floor, essentially as unconsolidated alluvium beneath the flood plains of larger streams tends to prevent erosion of underlying bedrock. As a result, fine-textured, integrated, low-gradient drainage systems with base levels close to the base of the soil cover are established locally at various altitudes. For a time the original polygonal networks are everywhere recognizable in the drainage patterns (Pl. 4, fig. 3), but, as the furrows are widened by continued erosion, the corners of the blocks become rounded and only their centers are left standing (Pl. 4, fig. 4) as systems of subcircular to subelliptical mounds.

Summary

The pimpled plains of eastern Oklahoma are evidently assignable to a category of surficial phenomena, sometimes called *Polygonboden* (Högbom, 1913–1914), which in some parts of the world includes features attributable to permafrost. The intermound furrow networks visible on airphotos of localities in eastern Oklahoma are comparable in pattern and coarseness of texture to the great polygonal networks of ice-filled fissures that commonly form in perennially frozen ground, but the eastern Oklahoma patterns bear an equally close resemblance to those of fissure networks caused elsewhere by desiccation and are comparable in some respects to intermound furrow patterns that appear to have originated, in some other parts of the United States, as a consequence of columnar jointing in the bedrock under the furrows. The data at hand offer little, if any, support to the possibility that the Oklahoma patterns are associated in origin with jointing in bedrock; they do, however, appear to establish desiccation as a possible agency in the origin of these patterns.

The transformation from systems of prismatic blocks enclosed by fissures to systems of mounds comparable to those of the pimpled plains may, apparently, be accomplished by one or more of three processes: (1) expansion of material that accumulates in the fissures; (2) subsidence of the ground along the fissures; and/or (3) erosion by rivulets which may form along the fissures. Assuming that the intermound furrow systems of eastern Oklahoma are related in origin to systems of fissure polygons due to desiccation, the mound relief there may be attributable to widening of grooves that have resulted from subsidence of the ground along the fissures of the polygonal networks.

Because the mound systems are present at various altitudes and occur locally on alluvial surfaces close to the present stream levels, the mound relief of eastern Oklahoma is probably not older than late Pleistocene.

* *Author's Note:* Since the foregoing discussion

was written, my attention has been drawn to an article by R. F. Black (*Polygonal patterns and ground conditions from aerial photographs*, Photogrammetric Eng., March 1952, p. 123–134) in which he cites T. T. Paterson's descriptions (*The effects of frost action and solifluction around Baffin Bay and in Cambridge district*, Geol. Soc. London, Quart. Jour., vol. 96, no. 381, 1940, p. 99–127) of polygonal fractures caused by seasonal frost and thaw in Subarctic localities outside the zone of permafrost. Black suggests that the polygons on Playa de Los Pinos, New Mexico, which Lang (1943) and Knechtel (1951) have attributed to desiccation, might be due to action of seasonal frost.

Though I am aware of no reason for excluding seasonal frost action from the possibilities relating to origin of the polygonal patterns associated with the pimpled plains of eastern Oklahoma, it is still my impression that the fracturing of Playa de los Pinos, in southwestern New Mexico at an altitude of about 4150 feet above sea level, has resulted from desiccation. While records of the U. S. Weather Bureau for comparable altitudes in southwestern New Mexico show that some winters are less mild than others, the duration of freezing temperatures at such altitudes is never long enough in that region to induce frost action of sufficient intensity to account for the large fractures that have formed on Playa de los Pinos. The records for Lordsburg, about 100 feet higher and only 10 miles to the east, indicate that the temperature in this neighborhood did not remain below the freezing point through all of any one day during the winter of 1949–1950; it rose above 50° F. on all except 12 days in that winter and reached 60° or higher on 14 days during December, the coldest month. When I visited the playa in late August of the following summer, after several weeks of severe drought, I saw many large open fractures which appeared to be quite fresh, as if they had formed as a result of drying after heavy rains that had fallen in July 1950, during which month the precipitation at Lordsburg was 2.03 inches, or approximately a quarter of an inch higher than normal. Lang (1943) reports that a resident of the neighborhood remembered seeing on the parched playa, during an "extreme but brief drought" in 1934, narrow, open fractures that "may have been two or three feet deep", which "were subsequently filled by collapse and inwash of surface material". When Lang visited the playa in 1937, it was largely covered by a sheet of water, and, when he returned in 1941, the polygonal fractures were visible only as broad, faint depressions, "more or less straight and trench-like, averaging 3 feet wide and perhaps an inch or more deep in the middle".

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U. S. Geological Survey, Washington, D. C.
Manuscript Received by the Secretary of the Society, May 29, 1951
Publication Authorized by the Director, U. S. Geological Survey

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 63, PP. 701-716, 4 FIGS. JULY 1952

STRONTIUM CONTENT OF LIMESTONES AND FOSSILS

By J. Laurence Kulp, Karl Turekian, and Donald W. Boyd

Abstract

Several hundred specimens of carbonate rocks and shells have been analyzed for strontium by a spectrographic method which gives a precision of 5 per cent or better. The average value of Sr/1000 Ca for all limestones was 0.71. Fossils are consistently higher but vary greatly, depending on the salinity and composition of the original ocean environment, the ratio of aragonite to calcite in the original shell, and the subsequent recrystallization history. On the average, fossils contain twice as much strontium as the surrounding carbonate matrix. Constancy of strontium content of a long range genus is evidence for similar marine environment at least since the Cretaceous. Temperature of the ocean water is a relatively insignifiant factor in determining the strontium content of fossils. No evidence could be detected for seasonal fluctuations in Sr/Ca ratio.

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Introduction

The geochemical data on the concentration of strontium in sedimentary rocks and carbonate fossils has been very limited until recently. Noll's work (1934), which consisted of the analysis of six limestones, nine fossils, and eight gypsum-anhydrite specimens, and the analyses of some corals by Tsuchiya (1944) were about all that was available until Odum (1950, 1951) made his comprehensive study of the biogeochemistry of strontium.

As part of a research program directed toward the evaluation of the ratio $Sr^{87/86}$ as a possible method of age determination of marine carbonates and shells, it was necessary to analyze a large number of representative carbonate rocks and shell fossils. In the course of this work, a rapid, high-precision spectrographic technique using a D.C. arc was developed.

With this technique in hand, it was desirable to analyze a large number of carefully selected specimens in an attempt to obtain significant geochemical and ecological data. Although this study of several hundred carbonate specimens was largely complete before Odum's work was known, there is little duplication. Odum (1950) was primarily interested in the worldwide distribution of strontium, so obtained only a moderate amount of data on natural marine carbonates. This research complements the previous work by adding considerably to the geochemical data on strontium in carbonates. Specifically, attempts to evaluate selectivity by various organisms, to study the relationship of fossil to matrix, and to obtain some general abundance data of strontium in natural marine carbonates and shells were the objects of this work.

Large parts of this research were initiated at the suggestion of Professor Norman D. Newell. His continuous interest, advice, and criticism were greatly appreciated. The suggestions on spectrographic technique by K. J. Murata and L. H. Ahrens were very helpful.

The authors wish to acknowledge the aid of numerous colleagues who kindly supplied valuable specimens. These include: Earl T. Apfel, R. L. Casanova, R. M. Dreyer, W. F. Evans, John Faick, J. B. Hadley, H. G. Hershey, L. Hintze, G. M. Kay, J. T. Lonsdale, Brian H. Mason, H. A. Rehder, H. B. Roberts, E. C. Stumm, and F. Stugard. The Indiana limestone specimens were procured through the Indiana Geological Survey.

Spectrographic Technique

The concentration of strontium in most carbonate rocks and fossils is too small for satisfactory chemical procedures. The spectrographic method is well suited to the concentration range (0.01 to 1.0% $SrCO_3$) and has the added advantage of great rapidity once the procedure is established. The method described below yields a standard deviation of less than 10 per cent for a single determination. Since each sample was run at least in duplicate and in some cases in triplicate, the analysis reported has a probable error of 3 to 5 per cent.

Odum (1950) did not obtain reasonable precision with the emission spectrograph in his preliminary experiments and turned to the flame photometer with which he developed a method which permitted satisfactory precision and rapidity. The emission spectrographic method used in this study achieved similar high precision and convenience. That the results reported here agree with those of Odum on the few similar samples provides a mutual check on the independent methods.

Since the method was to be used on samples of similar composition ($CaCO_3$) containing reasonably constant concentrations of calcium, a calcium arc line 4578.57 was used in the internal standard method with the strontium arc line 4607.331.

Eastman Kodak Spectrum Analysis No. 1 Film was used. Although the lines chosen lie in the range of a rapidly changing gamma, the measurements indicated that they are sufficiently close to each other that no appreciable error was encountered from this source.

The spectrograph was an Applied Research Laboratories 1.5 meter grating instrument. A.D.C. arc of 16.5 amperes was used, which decreased background and burning time. Platform-type electrodes were used with anode excitation. Time for complete burning was about one minute under these conditions. The presence of absorbed water, especially in the case of certain fossils, caused frequent popping. This was eliminated by pre-arcing the sample for 6 seconds at 7 amperes. D-19 developer was used for 3 minutes at 70° F.

The preparation of the working curve was carried out with synthetic $CaCO_3$–$SrCO_3$ standards. A problem developed at this stage because of the high concentration of strontium (0.1 per cent) in analytical (C.P.) $CaCO_3$. By a process of three consecutive extractions of the C.P. $CaCO_3$ with hot iso-amyl alcohol, the $SrCO_3$ content was cut to 0.02 per cent. Further correction was made graphically as described in the literature.

Self-absorption begins to be visible in Sr 4607 above 0.1% $SrCO_3$. Attempts to increase the upper range failed, leaving no alternative but to dilute samples which on the first determination proved to have a $SrCO_3$ concentration greater than 0.1 per cent.

Where dilution was necessary and the non-

carbonate impurities were noticeably high, the sample was first leached and reprecipitated to guarantee an essentially $CaCO_3$ matrix. However, the Sr-Ca line intensity ratio appeared independent of Silica content to at least 30% SiO_2.

TABLE 1.—STRONTIUM CONTENT OF CERTAIN INDIANA LIMESTONES*

Geologic Age	Formation	$\frac{Sr}{1000\ Ca}$	%$MgCO_3$†
Pennsylvanian, Monongahela	Maria Creek	0.92 Av. 0.72	6.44
Pennsylvanian, Pottsville	Minshall	0.73	2.63
Mississippian, Chester	Glen Dean	0.58	5.47
	Beech Creek	0.55	1.86
	Beaver Bend	0.49	0.96
	Paoli	0.55	0.80
	Aux Vases	0.65 Av. 0.46	2.39
Mississippian, Meramec	Ste. Genevieve (1)	0.36	0.98
	Ste. Genevieve (2)	0.30	0.91
	St. Louis	0.39	5.66
	Salem	0.29	1.06
Devonian, Hamilton	Beechwood	1.87	14.2
	Silver Creek	3.90	18.9
	Speeds	3.44 Av. 2.48	5.03
Devonian, Onondaga	Jeffersonville	2.25	4.45
Devonian, Oriskany	Geneva	1.03	38.6
Silurian, Cayugan	Kenneth	1.74	1.04
	Kokomo	1.82	12.8
Silurian, Niagaran	Huntington	1.08	41.1
	Louisville	1.69 Av. 1.65	13.9
	Liston Creek	1.08	4.76
	Laurel	1.51	28.2
	Mississinewa	2.66	23.1
	Osgood	1.58	16.4
Silurian, Albion	Briarsfield	0.82	1.09
Ordovician, Cincinnatian	Elkham	0.83	12.2
Ordovician, Cincinnatian	Whitewater	1.69	3.34
Ordovician, Cincinnatian	Saluda	0.61	2.82

* Complete locality data can be obtained from the authors.

† The $MgCO_3$ determinations are taken from Report of Progress No. 3, *Crushed Stone in Indiana*, By J. B. Patton as revised by R. K. Leininger, Spectrographer, Indiana Geological Survey.

RESULTS

Limestones, Dolomites, and Marbles

Table 1 shows the results for some of the Indiana limestones. The difference between the Mississippian and lower limestones is striking. (It should be emphasized that the probable error is about 5 per cent for all of these determinations.) This may suggest a consistent difference in salinity. Although the Mississippian limestones are all low in magnesium and many of the Silurian and Devonian limestones are high in magnesium, the correlation of strontium and magnesium content is only fair. Since high magnesium content is commonly but not necessarily an effect of increased salinity, this lack of complete correlation is not surprising.

For various samples taken 144 miles along

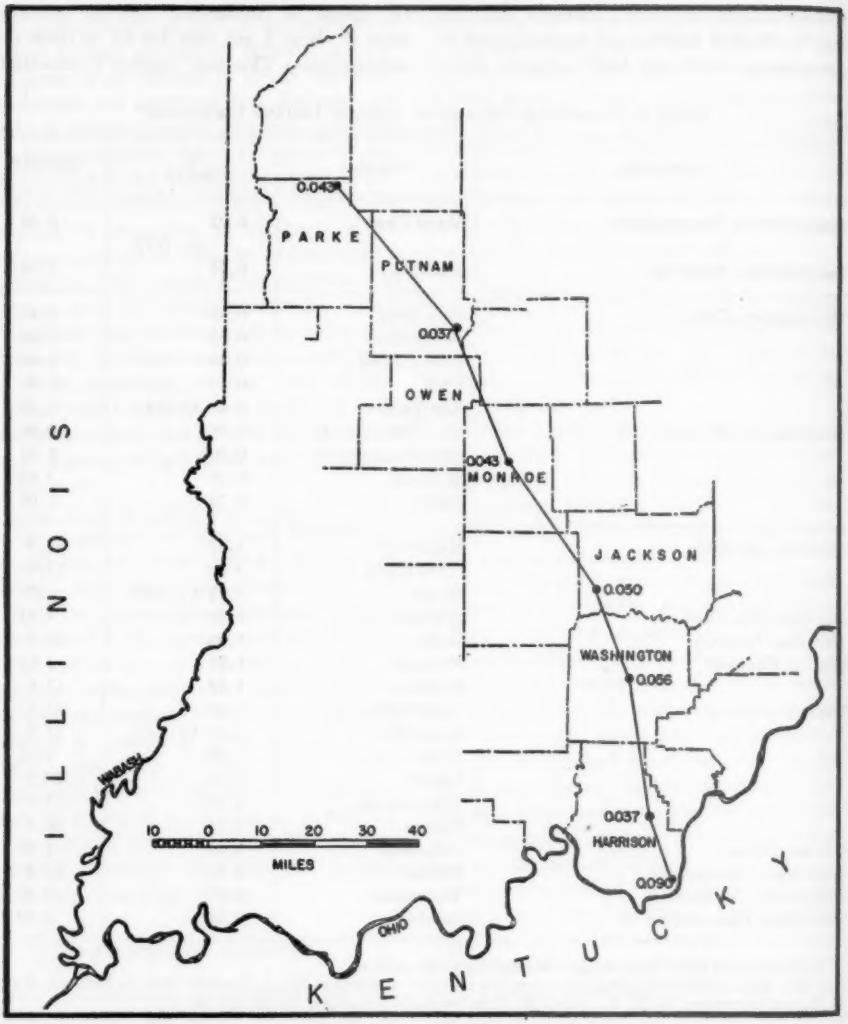

FIGURE 1.—STRONTIUM CONTENT ALONG STRIKE OF THE HARRODSBURG FORMATION IN SOUTHEAST INDIANA

TABLE 2.—STRONTIUM CONTENT OF HARRODSBURG LIMESTONE (MISS. OSAGE)

Location		$\frac{Sr}{1000\ Ca}$
Parke County	NW ¼ SE ¼ S 7, T17N, R6W	0.29
Putnam County	SW ¼ S 23, T13N, R3W	0.25
Monroe County	SW ¼ NW ¼ S28, T9N, R1W	0.29
Jackson County	SE ¼ SE ¼ S29, T5N, R3E	0.34
Washington County	SE ¼ NW ¼ S20, T2N, R4E	0.38
Harrison County (N)	NE ¼ SW ¼ S25, T3S, R4E	0.25
Harrison County (S)	SW ¼ NE ¼ S27, T5S, R5E	0.61

TABLE 3.—STRONTIUM CONTENT OF VARIOUS CARBONATE ROCKS

Age	Location and description	$\frac{Sr}{1000\ Ca}$
Precambrian	Sprout Brook, N. Y., limestone	0.17
	Inwood, New York City, marble	0.41
	Port Henry, N. Y., marble	0.24
	East of Lalaboogie Lake, Ont., Grenville marble	0.62
	Adirondacks, Grenville marble	1.33
	Glacier Park, Mont., Belt Series limestone	0.16
	Menominee Range, dolomite	0.25
	Franklin, N. J., Franklin limestone	2.24
	Middletown, Conn., marble in Bolton schist (1)	2.45
	Middletown, Conn., marble in Bolton schist (2)	2.04
Cambrian	DuNoir, Wyo., 25 feet below top Gallatin formation	0.25
Ordovician	Tamworth, Ont., Black River	0.52
	Fort McCord, Penna., Doylesburg member Shippensburg limestone	0.41
	Burnt River, Ont., Black River	0.33
	East Martinsburg, N. Y., Black River	0.43
	Medante, Ont., Black River	0.37
	Perry County, Texas, Zell member, Macy formation (1)	1.22
	Perry County, Texas, Zell member, Macy formation (2)	0.70
	Perry County, Texas, Zell member, Macy formation (3)	1.33
	Perry County, Texas, Zell member, Hager formation (1)	0.88
	Perry County, Texas, Zell member, Hager formation (2)	0.90
Silurian	Ferry Island, N. Y. High Decker formation	1.19
Devonian	Schoharie, N. Y. Manlius formation	0.56
	Traverse Group (Mich.), Kilban limestone	0.47
	Anderson limestone (Mich.) coral bed	0.13
	(Mich.) core rock of bioherm	0.12
	(Mich.) Stromatoporoid layer	0.47
	Rogers City limestone	0.21
	Charlevoix limestone	0.15
	Dundee limestone	0.24
	Kagaskewing, Mich.	0.11
	Alpena, Mich., (1) Middle Devonian	0.17
	Alpena, Mich., (2) Middle Devonian	0.44
	Detroit, Mich., salt shaft Middle Devonian	0.18
	Black Lake, Mich.	0.20
	Formosa, Ont., Onondaga	0.25
	Schoharie, N. Y., Rondout waterlime	0.71
Mississippian	Bedford, Ind., Indiana limestone formation	0.25
	Platte Canyon, Wyo., Madison limestone	0.12
	Batesville, Arkansas, Spring Creek formation	0.77
	Black Hills, S. D., Pahasapa limestone	0.13
	Austwick, Yorkshire, Porcellaneous limestone	0.26
	Hanover, N. M., Lower Blue limestone	1.73
	Hanover, N. M., Crinoidal limestone	0.14

TABLE 3.—CONTINUED

Age	Location and description	Sr / 1000 Ca
Pennsylvanian	Hanover, N. M., Upper Blue (1)	0.43
	Hanover, N. M., Upper Blue (2)	0.40
	Kansas, Leavenworth limestone	1.30
	Plumbago Canyon, Wyo., Casper formation	1.22
Permian	Grand Canyon, Toroweap limestone	0.18
	Kazan, Tartaria, Kazanian	0.24
	Muley twist, Ariz., Kaibab limestone	0.67
	Kansas, Permian limestone	0.31
Triassic	Freezeout Hills, Wyo., Alcova	0.39
	Shasta County, Calif., Ben Mt.	1.46
	Arizona Canyon, Nev., Star Peak limestone	3.12
Jurassic	Dubois, Wyo., Lower Sundance	0.45
Cretaceous	Fort Stockton, Pecos County, Texas, Duck Creek limestone	0.47
	Edwards County, Texas, Edward limestone	0.27
	Comanchean limestone	0.29
	Wyo., Pierre formation	1.33
	Tamosopo Canyon, San Luis Potosi, Mexico, Tamosopo Series	0.37
Tertiary	Oligocene-Miocene, Vaquerous formation	1.09
	Aruba Bon Ayre, Curacao, Miocene	1.50

Deep-Sea Globigerina Ooze Specimens

Age	Depth in Core	Depth in Water	Latitude	Longitude	Sr/1000 Ca
A167-21 Upper Eocene	200 cm.	1454 m	29°49′N	76°35′W	12 ± 3
A167-22 Oligocene	190 cm.	2204 m	29°49.8′N	76°28½′W	10 ± 3
A167-29 Late Miocene	400 cm.	1829 m	28°26½′N	76°40′W	10 ± 3

the strike (Fig. 1) of the Harrodsburg limestone in Indiana (Table 2), the constancy of Sr/Ca ratio reflects homogeneous conditions of sedimentation. Only the southern-most sample lies appreciably outside the experimental error of the mean.

Table 3 lists the strontium data for the other rocks tested. There is no trend with age. The total spread of the Sr/Ca ratio is quite large since all types of limestones, both recrystallized and primary, are represented. Within a narrow stratigraphic interval and small area, however, the strontium concentration is quite constant. See for example, the Ordovician Black River, the Perry County, or the Michigan Devonian specimens.

The Harrodsburg and Tamworth limestones are known to contain celestite. The strontium content of these limestones is probably low due to recrystallization during which some of the strontium has been liberated. When a limestone has had such a history, the present Sr/Ca ratio cannot be used to define the salinity of the ocean from which the limestone was deposited.

The deep-sea ooze samples show high values of Sr/Ca.

Fossils

The strontium content of a randomly selected group of fossil types is shown in Table 4. These

TABLE 4.—STRONTIUM CONTENT OF RANDOMLY SELECTED FOSSILS

Fossil	Age	Formation	Sr/1000 Ca
Brachiopods			
Pentamerus oblongus	Silurian	Clinton limestone (Rochester)	0.44
Atrypa reticularis	Silurian	Clinton limestone	0.36
Spirifer radiatus	Silurian	Waldron, Ind.	0.75
Spirifer hungerfordi	Devonian	Lime Creek, Iowa	0.61 Av. 0.66
Athyris spiriferoides	Devonian	Hamilton, N. Y.	0.78
Terebratula	Triassic	Vienna, Austria	0.78
Lingulepis	Cambrian	St. Croix	4.75*
Crinoids			
Caryocrinites ornatus	Silurian	Clinton	0.51
Eucalyptocrinus coelatus	Silurian	Waldron, Ind.	0.80 Av. 0.45
Actinocrinus aequalis	Mississippian	Burlington, Iowa	0.31
Crinoid stems	Pennsylvanian	Brownwood, Texas	0.20
Corals			
Pachyphyllum woodmani	Devonian	Lime Creek, Iowa	0.26
Stromatopora	Devonian	Hamilton, N. Y.	0.13
Caninia torquia	Pennsylvanian	Stonewall Quadrangle, Okla.	0.47 Av 0.22
Coral bed	?	Anderson limestone	0.14
Core rock of bioherm	?	Anderson limestone	0.12
Lophophyllidium proliferum	Pennsylvanian (matrix?)	Thrifty formation	1.62
Gastropods			
Bellerophon crassus	Pennsylvanian	Holdenville formation	0.12
Platyceras aequilaterale	Mississippian	Keokuk	0.16
Pelecypods			
Cardium	Upper Miocene	Yorktown (Pa.?)	2.95
Ostrea	Eocene	Midway	0.81
Belemnites			
Belemnites elongatus	Jurassic	Germany	0.90
Belemnites densus	Jurassic	Sundance	1.00
Belemnite	Cretaceous	Atlantic Highlands	0.87
		Navesink Marl	0.99
Algae			
Algal reef	Belt Series Proterozoic		0.14
Algal matrix	Belt Series Proterozoic		1.80
Echinoderms			
Dendraster diegoensis	Middle Pliocene	San Diego, Calif.	0.21
			0.21
			0.22

* Chemical tests showed high phosphate content.

samples were some of the first groups of analyses. At this stage the primary interest was to determine whether a certain fossiliferous horizon had sufficient strontium for practical separation. Hence, no great effort was taken to separate the shell and limestone matrix. Therefore in certain fossil types, such as the corals, the strontium content is probably not representative of the fossil itself. The high value of *Lingulepsis* is probably related to the high phosphatic content of the shell as evidenced by chemical tests.

No trends with age or fossil type appear. Note the constant value for the *Dendraster*

TABLE 5.—STRONTIUM CONTENT OF FOSSILS FROM THE CALVERT FORMATION (MIOCENE), PLUM POINT, MARYLAND

Genus	Species	Sample	Sr/1000 Ca
Phylum Mollusca Class Gastropoda			
1 *Scaphella*	*solitaria* (Conrad)	a	3.3
		b	3.3 Av. 3.21
		c	3.0
2 *Calyptrea*	*aperta* (Solander)	a	2.7
		b	3.3 Av. 2.93
		c	2.9
3 *Xenophora*	*conchyliophora* (Born)	a	2.7
		b	2.9 Av. 2.80
		c	2.9
4 *Crepidula*	*fornicata* (Linnaeus)	a	2.8
		b	2.4 Av. 2.80
		c	3.1
5 *Polinices*	*heros* (Say)	a	2.3
		b	2.6 Av. 2.39
		c	2.3
6 *Crucibulum*	*costatum* (Say)	a	2.8
		b	2.8 Av. 2.80
		c	2.8
7. *Turritella*	*variabilis* var. *exaltata* (Conrad)	a	2.9
		b	3.5 Av. 3.00
		c	2.7
8 *Niso*	*lineata* (Conrad)	a	2.3
		b	2.3 Av. 2.53
		c	2.9
9 *Drillia*	*limatula* (Conrad)	a	2.1
		b	2.7 Av. 2.46
		c	2.6
10 *Epitonium*	*pachypleura* (Conrad)	a	1.8
		b	2.4 Av. 2.12
		c	2.1
11 *Pleurotoma*	*bellacrenata* (Conrad)	a	3.0
		b	2.6 Av. 2.86
		c	3.0
12 *Calliostoma*	*bellum* (Conrad)	a	2.1
		b	1.9 Av. 2.25
		c	2.7
Phylum Mollusca Class Pelecypoda			
13 *Crassatellites*	*melinus* (Conrad)	a	7.4
		b	6.8 Av. 6.57
		c	5.6
14 *Corbula*	*idonea* (Conrad)	a	3.7
		b	3.0 Av. 2.93
		c	2.2
15 *Cardium*	*leptopleurum* (Conrad)	a	3.5
		b	3.3 Av. 3.28
		c	3.1

TABLE 5.—CONTINUED

Genus	Species	Sample	Sr/1000 Ca
Phylum Mollusca Class Gastropoda			
16 *Dosinia*	*acetabulum* (Conrad)	a	2.2
		b	2.9 Av. 2.83
		c	3.4
17 *Venus*	*mercenaria* (Linnaeus)	a	3.2
		b	3.1 Av. 3.00
		c	2.8
18 *Arca*	*subrostrata* (Conrad)	a	2.8
		b	3.3 Av. 2.73
		c	2.3
19 *Pecten*	*madisonius* (Say)	a	1.9
		b	3.6 Av. 2.05
		c	1.6
20 *Glycymeris*	*parilis* (Conrad)	a	3.0
		b	2.9 Av. 2.73
		c	2.2
21 *Nuculana*	*liciata* (Conrad)	a	3.1
		b	2.6 Av. 2.59
		c	2.2
Phylum Mollusca Class Scaphopoda			
22 *Dentalium*	*attenuatum* (Say)	a	3.8
		b	3.9 Av. 3.88
		c	3.9
annelid tube		a	12.0

specimens taken from a single horizon in a small area.

In order to assess the selectivity of various fossils for strontium, a large number of specimens (Table 5) were taken from one horizon, the Miocene Calvert formation at the outcrop near Plum Point, Maryland. All the genera sampled are represented today by species living in the waters along the Florida coast, the West Indies, and/or the Gulf of Mexico. There are 23 species representing 23 genera. Three specimens of each species were analyzed, each one in duplicate or triplicate.

It appears that, under these conditions, the strontium content of a particular genus is relatively constant. Further, except for certain individuals, the strontium concentration is quite constant within a class. For a given environment, most fossils appear to build the same Sr/Ca ratio into their shells. There is evidence from this table, however, that certain animals are selective. For example, the genera *Crassatellites* and *Dentalium* show strontium enrichment. The genera *Pecten* and *Epitonium*, on the other hand, are slightly, but significantly, lower in strontium than the others. The worm tube is very much higher in Sr/Ca. To check the unique behavior of worm tubes, two modern representatives were analyzed semi-quantitatively. The result showed them to contain a Sr/Ca ratio of about 10.

Bøggild (1930) makes several observations that may help to understand the high value of the Sr/Ca ratio of *Crassatellites* and *Dentalium* and the low value of *Pecten*. *Pecten* is described (p. 265) as having an essentially calcitic shell with an aragonite layer interbedded between the inner and outer calcitic layers. This layer is "rather thin" and not easily visible. Since calcite has a lattice that is less amenable to Sr than the aragonite lattice, this might explain the low Sr content in this shell. *Pecten*

shells (*Chlamys* P11, P12) in Table 10 are also lower in Sr than other modern shells of the same phylum.

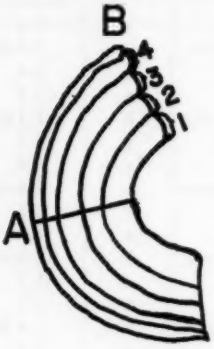

FIGURE 2.—SECTION OF *Atractites pusillus* HAUER

TABLE 6.—STRONTIUM CONTENT

Atractites Zones (Fig. 2)

Zone	Section	Sr/1000 Ca	Average
1	A	1.8	1.8
1	B	1.9	
2	A	2.1	2.1
			Mean 2.1
2	B	2.1	
3	A	1.7	2.0
3	B	2.3	
4	A	2.5	2.4
4	B	2.3	
Glycymeris Zones (Fig. 3)			
1	A	1.9	
2	A	1.6	
3	A	1.6	
4	A	1.8	

Crassatellites is described (p. 275, 279) as essentially an aragonitic shell with calcite very rare. This is generally true of the order Homomyaria.

The Scaphopoda (p. 240) are listed as having aragonitic shells. This seems consistent with the high recorded Sr value for *Dentalium*.

Initially most molluscs have a mixture of calcite and aragonite in their shells. This may largely explain the intermediate values of the other samples. It should be emphasized that these fossils from the Calvert formation have

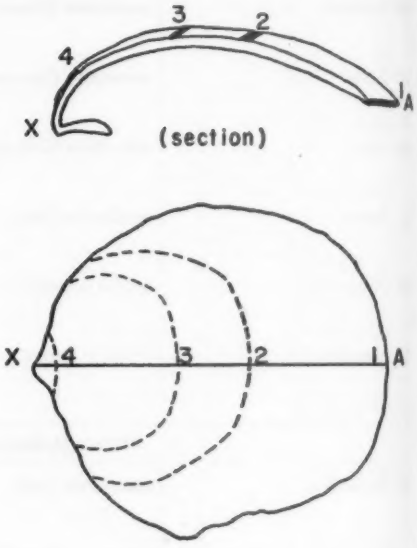

FIGURE 3.—SKETCH OF *Glycymeris rubicus* LAMARCK

not been recrystallized. If fossils are recrystallized, these relationships would be obscured.

An average value for all these shells excluding the five special cases mentioned above, is 2.93 Sr/1000 Ca with a mean deviation of 0.24, which is only slightly larger than the experimental error.

Another type of study was conducted on two individuals to see if annual or life-span trends in strontium content occur. The first specimen (Fig. 2), was *Atractites pusillus* Hauer, a belemnite from the Upper Muschelkalk of Haliluci, Bosnia. Samples were taken along the sections A and B (Table 6). Averages are just outside the probable error (about ± 0.1 here) of the mean. They may suggest slight annual oscillation or they may be fortuitous. A sampling cut (Fig. 3) across a shell of *Glycymeris rubicus* Lamarck from New Zealand (recent) gave essentially the same result (Table 6). There seems to be no appreciable variation with age.

TABLE 7.—STRONTIUM CONTENT OF BRACHIOPODS AND CARBONATE MATRIX* IN VARIOUS HORIZONS

Description	Sr/1000 Ca	
	Sample	Matrix
1 *Martinia rhomboidalis* Girty; Carlsbad limestone, Permian, SE New Mexico (Whiteman loc. 725)	as 0.53 bs 0.58 Av. 0.56 cs 0.57	am′0.11 bm′0.26 Av. 0.20 cm′0.23
2 *Rhynchotrema minnesotense* Sardeson; R. Decorah formation, Middle Ordovician, Wisconsin	as 0.55 Av. 0.55	am 0.27 Av. 0.27
3 *Strophomena rugosa* Blainville; Cincinnatian series, Upper Ordovician, Cincinnati, Ohio	as 0.64 bs 0.88 Av. 0.75 cs 0.64	 bm′0.48 Av. 0.61 cm′0.71
4 *Rhipidomella oblata* Hall; Lower Helderbergian stage, Devonian, New York	as 1.73 bs 1.93 Av. 1.83 cs 1.83	am′1.25 bm′1.36 Av. 1.29 cm′1.15
5 *Orthostrophia strophomenoides* Hall; Haragan shale, Helderbergian stage, Devonian, Murray County, Okla.	as 1.37 bs 1.37 Av. 1.36 cs 1.22	am 0.40 bm 0.33 Av. 0.38 cm 0.39
6 *Atrypa reticularis* (Linnaeus); Cedar Valley limestone, Upper Erian series, Devonian, Solon, Iowa	as 0.55 bs 0.50 Av. 0.51 cs 0.50	am 0.17 bm 0.18 Av. 0.16 cm 0.14 bm′0.27 Av. 0.26 cm′0.24
7 *Leptaena rhomboidalis*; New Scotland formation, Helderbergian stage, Devonian, Schoharie, N. Y.	as 2.10 bs 1.36 Av. 1.36 cs 0.58	am′1.42 bm′1.36 Av. 1.02 cm′0.20
8 *Mucrospirifer thedfordensis* Shimer and Grabau; Upper Hamilton formation, Erian series, Devonian; Thedford, Ont.	as 1.07 bs 1.00 Av. 1.02 cs 0.87	am′0.39 bm′0.52 Av. 0.48 cm′0.54
9 *Composita subtilita*; Upper Boggy shale, Des Moinesian series, Pennsylvanian, Stonewall quadrangle, Okla.	as 0.78 bs 0.51 Av. 0.61 cs 0.44	am 0.39 bm 0.52 Av. 0.48 cm 0.62
10 *Marginifera muricata* (Norwood and Pratten); Lower Boggy shale, Des Moinesian series, Pennsylvanian, Stonewall quadrangle, Okla.	as 1.80 bs 0.82 Av. 1.57	am′1.73 bm 1.32 Av. 1.49
11 *Leiorhynchus rockymontanus* (Marcou); Upper Wetumka shale, Des Moinesian series, Pennsylvanian, Stonewall quadrangle, Okla.	as 0.44 cs 0.33 Av. 0.41	am 0.57 cm 0.44 Av. 0.48
12 *Derbyia crassa* (Meek and Hayden); Upper McAlester formation, Upper Lampasian series, Pennsylvanian, Stonewall quadrangle, Okla.	as 0.95 bs 0.59 Av. 0.82 cs 0.88	am′0.52 bm′0.36 Av. 0.48
13 *Atrypa reticularis* (Linnaeus); Rochester shale formation, Clinton group, Silurian, Lockport, N. Y.	as 1.26 Av. 1.50 cs 1.77	am′0.54 Av. 0.48 cm′0.44

TABLE 7.—CONTINUED

Description	Sr/1000 Ca	
	Sample	Matrix
14 *Atrypa reticularis* (Linnaeus); Clinton group, Niagaran series, Madison, Ind.	as 1.94 bs 1.67 Av. 1.97 cs 2.24	am 0.19 bm 0.24 Av. 0.20 cm 0.27
15 *Homoeospira evax* (Hall); Waldron Shale, Upper Niagaran series, Silurian, Indiana	as 0.58 bs 0.55 Av. 0.61 cs 0.68	am 0.12 bm 0.24 Av. 0.17 cm 0.22

* Matrix taken from inside of shell (m); from outside (m′).

TABLE 8.—STRONTIUM CONTENT OF FOSSILS AND MATRIX FROM LOWER FRANCIS FORMATION, MISSOURIAN SERIES, PENNSYLVANIAN, STONEWALL QUADRANGLE, OKLAHOMA

Description	Sr/1000 Ca	
	Sample	Matrix*
Phylum Brachiopoda		
Composita subtilita (Shepard) (matrix inside)	as 0.50 bs 0.57 Av. 0.52 cs 0.50	am 0.59 bm 0.46 Av. 0.50 cm 0.44
Chonetes granulifer Owen	as 1.06 bs 0.88 Av. 0.90 cs 0.75	am 0.31 Av. 0.36 cm 0.41
Derbyia crassa (Meek and Hayden)	as 0.71 bs 0.71 Av. 0.84 cs 1.09	am 0.35 bm 0.32 Av. 0.35 cm 0.38
Dictyoclostus "semireticulatus" (Martin)	as bs 1.16 Av. 0.97 cs 1.02	am 1.64 bm 1.87 Av. 1.75
Neospirifer cameratus (Morton)	as 0.58 bs Av. 0.70 cs 0.61	am 0.66 bm 0.58 Av. 0.69 cm 0.82
Linoproductus cora (D'Orbigny)	as 1.23 Av. 1.20 cs 1.19	am 1.02 Av. 1.10 cm 1.19
Phylum Mollusca Class Gastropoda		
Worthenia cf. *tabulata* (Conrad) (matrix partly inside and partly outside)	as bs 0.10 Av. 0.10 cs 0.11	am 0.13 bm 0.20 Av. 0.18 cm 0.22

* Matrix is from outside the shell unless noted differently.

TABLE 9.—STRONTIUM CONTENT OF SPECIMENS OF A LONG-RANGING GENUS

Description	Sample	Sr/1000 Ca
1 *Turritella coalvillensis* Meek; Cretaceous, Utah	a	1.1
	b	2.0 Av. 1.67
	c	1.8
2 *Turritella mortoni* Conrad; Aquia formation, Lower Eocene, Md.	a	2.2
	b	1.6 Av. 2.01
	c	2.2
3 *Turritella carinata*; M. Eocene, Claiborne, Ala.	a	1.6
	b	1.7 Av. 1.60
	c	1.5
4 *Turritella* sp. Eocene; Whitlock, Texas	a	5.1
	b	3.2 Av. 4.15
	c	4.2
5 *Turritella gatunensis* Conrad; Gatun formation, Lower Miocene, Gatun Lock, Panama	a	2.3
	b	1.6 Av. 1.69
	c	1.2
6 *Turritella indenta* Conrad; Calvert formation, Lower Miocene, Plum Point, Md.	a	2.2
	b	1.5 Av. 1.72
	c	1.4
7 *Turritella subgranulifera*; Upper Miocene; York River, 4 miles above Yorktown, Va.	a	2.9
	b	2.6 Av. 2.58
	c	2.2
8 *Turritella jewetti* Carpenter; Pliocene; Dead Mans Isle, Calif. (only 2 indiv.)	a	2.5
	b	2.5 Av. 2.48
9 *Turritella annulata* Kien; Recent; Gambia	a	2.5
	b	2.7 Av. 2.52
	c	2.4

To investigate the relationship of the strontium content of a shell to that of a carbonate matrix, brachiopods and surrounding matrices for 15 horizons were analyzed (Table 7). Where available, three separate samples of fossils and three of matrix were run for each horizon. In each case the matrix adhering to the shell was chosen. The most striking result is the higher strontium content of the fossil over the carbonate matrix. These relationships are independent of the total strontium present. The strontium content of the matrix is possibly determined by the ratio of organic to inorganic carbonate and the shell types represented in the organic fraction. Where the fossil and matrix have the same strontium concentration, the matrix is probably mainly organic and made up largely of shell fragments of that class of fossils or other classes with similar strontium content.

TABLE 10.—STRONTIUM CONTENT OF RECENT SHELLS FROM COLD AND WARM WATER ENVIRONMENTS

Sample	Description	Sr/1000 Ca
P I 1a	*Chlamys islandicus* Muller, Hudson Bay, Canada	1.3
b		1.1 Av. 1.2
P I 2a	*Chlamys sentis* Linne, Tortugas, Fla., 15 fathoms	0.9
b		0.9 Av. 0.9
P II 1a	*Crepidula fornicata* Say, Massachusetts Bay	2.7
b		1.7 Av. 2.2
P II 2a	*Crepidula fornicata* Say, Sarasota Bay, Fla.	1.9
b		2.5 Av. 2.2
P III 1a	*Cerastoderma ciliatum* Fabricus, off Cape Cod, 26 fathoms 44°F	2.4 Av. 2.4
P III 2a	*Trachycardium egmontianum* Shuttleworth, Tampa Bay, Fla.	1.9 Av. 1.9

The average of the individual species average compared with the corresponding matrix average is as follows:

average of species: 1.03 Sr/1000Ca

average of matrices: 0.51 Sr/1000Ca

It appears from this random study of fossils in matrix that the animal concentrates Sr relative to its matrix (inorganic Ca $CaCO_3$?) in the ratio of about 2:1. This may be related to

the original aragonite/calcite ratio in the two carbonates. It is possible, however, that the finer matrix material is more readily replaced or recrystallized, liberating strontium and thus lowering the Sr/Ca ratio.

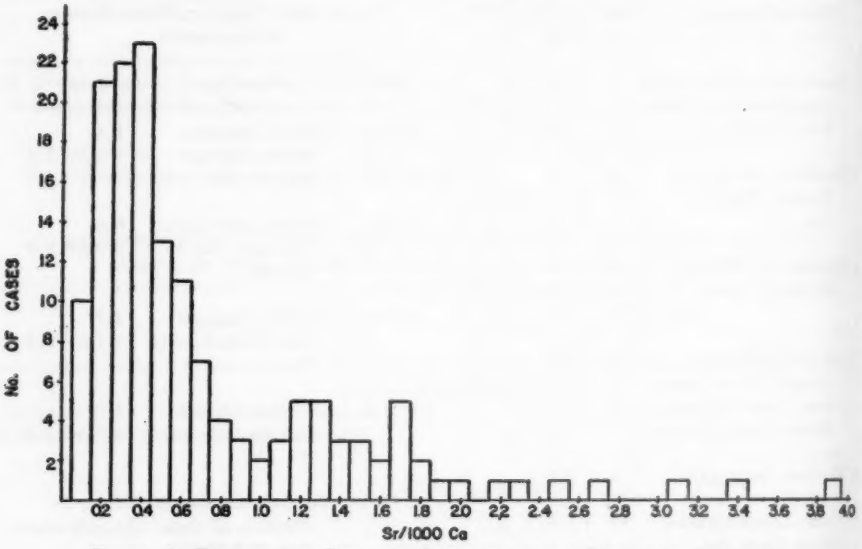

FIGURE 4.—DISTRIBUTION CURVE OF STRONTIUM IN THE LIMESTONES ANALYZED

A study was attempted to choose fossils and matrices from a common horizon (Table 8). Unfortunately, a suite was readily available only from the Francis formation, and this formation is known to be variable lithologically (Morgan, 1924).

However, some interesting information can be obtained from the data. The brachiopods analyzed were similar in Sr concentration despite the variable concentration of the matrix. The Sr/Ca of the matrix may vary for several reasons as discussed above.

No explanation is apparent for the very low value of the one gastropod available for study (0.10 Sr/1000Ca).

The strontium content of a series of non-recrystallized specimens of a long-ranging genus, *Turritella*, was studied. The average strontium content is quite constant with time (Table 9). This is consistent with Odum's results (1950) on several other animal types. The only anomalous value is from a species collected in the Eocene of Texas.

To examine the possible effect of temperature on the incorporation of strontium in a shell, living forms with representatives in both warm and cold water were analyzed (Table 10). The data suggest that there is no significant temperature effect for these fossil types under the existing salinity conditions. Odum (1950) also found no variation in Sr/Ca ratio in *Polynices* and *Lithothamnium* specimens from cold and warm water.

DISCUSSION

The number of specimens analyzed in this study is sufficient to indicate probable ranges of the strontium content of carbonate rocks and fossils. Figure 4 shows the strontium distribution in all of the limestones analyzed in this study, including limestone matrix samples from Tables 7 and 8. Apparently strontium enrichment may occur over a considerable range but the Sr/Ca lower limit has a sharp cutoff. The average value for all samples of the ratio Sr/1000Ca is 0.71. If the matrix samples which

have some shale content are omitted the ratio is 0.80.

These results are summarized and compared with those of Odum (1950) in Table 11. Where similar material has been analyzed, the results are in rather close agreement. The gross averages for fossils are not comparable since they were selected differently. The values in this study are more likely representative of the actual distribution in the fossils of the earth's crust. Odum's values are weighted heavily on the side of aragonitic, nonrecrystallized material. The difference in the limestone averages is due in part to the smaller number used in Odum's work. Further it is not clear how much unrecrystallized material is included in his average. On the whole the agreement is very good considering the fact that entirely different sets of samples were used and were analyzed by different techniques. The geochemical relation in marine carbonates are, therefore, much more closely defined than heretofore.

The wide geologic and geographic distribution represented in this study gives assurance that large differences in the averages are significant. Thus it is clear that limestones are considerably lower in strontium than most common fossils.

A number of workers (summary by Odum, 1950) have shown that large quantities of strontium can be taken up in shells of animals grown under controlled conditions and that the relationship between Sr/Ca in the shell is almost directly related to the ratio in the environment.

Thus the primary factor which determines the Sr/Ca ratio in the shell or limestone is the Sr/Ca of the water from which these are deposited. The Sr/Ca of the water in turn is related to the salinity (temperature is of very minor importance) and the source. The secondary factor is the crystal lattice being formed. Other things being equal, strontium is admitted more readily to aragonite than calcite. In organisms the ratio of the growing shell is apparently independent of the age of the animal, the season of the year and the biological processes. However, the ratio is strongly dependent on the type of crystal lattice which the organism prefers. Finally recrystallization always lowers the Sr/Ca ratio in varying amounts so that, after recrystallization, the strontium content can not be used as an index of initial conditions.

TABLE 11.—SUMMARY OF THE STRONTIUM ABUNDANCE DATA IN CARBONATES

Description		Sr/1000 Ca
Average of 155 Limestones (This Study)		0.71
Average of 40 Limestones (Odum)		1.21
Range 78 Paleozoic Limestones (This Study)		0.11–3.90
Average 78 Paleozoic Limestones (This Study)		0.76
Range (25) Paleozoic Limestones (Odum)		0.3 to 1.4
Average (25) Paleozoic Limestones (Odum)		0.9
Average of 155 fossils Calvert formation (Miocene)		3.10
Average 103 unrecrystallized fossils (Brachiopods, Pelecypods, Gastropods)		2.76
Average (198) all fossils studied (includes many recrystallized)		1.88
	Cases	
Miocene *Venus*	3 (This Study)	3.0
	2 (Odum)	3.1
Pecten	3 (This Study)	2.1
	2 (Odum)	2.5
Dentalium	3 (This Study)	3.8
	1 (Odum)	2.4

It is significant that all of these carbonate specimens contain considerably more than 0.01% $SrCO_3$, and that the average is closer to 0.1% $SrCO_3$. This appears to be high enough for practical separation of a few milligrams of strontium from such materials for isotopic analyses. Thus if the strontium isotope ratio method of age determination (Wickman, 1948) proves valid, essentially all carbonate rocks and fossils are suitable for measurement.

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LAMONT GEOLOGICAL OBSERVATORY, PALISADES, NEW YORK.

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, NOVEMBER 9, 1951.

PROJECT GRANT #549-49.

LAMONT GEOLOGICAL OBSERVATORY CONTRIBUTION No. 59.

EXPLANATION

Sedimentary Rocks

Recent	Qd	Recent sand dunes	QUATERNARY
	Qa	Recent alluvium	
	Qb	Lake Bonneville sediments	
Pliocene? Pleistocene	Qpb	Pre Bonneville sediments	
Oligocene?	Tf	Fool Creek conglomerate	TERTIARY
Late Cretaceous and Paleocene	KTn	North Horn (?) formation	CRETACEOUS TERTIARY AND TERTIARY
Upper Cretaceous	Ki	Indianola group(?)	CRETACEOUS TERTIARY
Upper Cambrian and Carboniferous	C	Carboniferous, undifferentiated	
Upper Cambrian and Ordovician (?)	€ls	Undifferentiated limestone and dolomite	CAMBRIAN AND ORDOVICIAN (?)

Upper Cambrian	€o	Ophir formation	CAMBRIAN AND ORDOVICIAN (?)
Lower Cambrian	€t	Tintic quartzite	
Upper Precambrian	pCp	Proterozoic quartzite, shale and limestone	PRECAMBRIAN (?)

Igneous Rocks

	Tv	TERTIARY

- 70 — Dip and strike of beds
- 80 — Dip and strike of overturned beds
- Vertical beds
- Bonneville shoreline
- Thrust fault
- Fault
- Syncline
- Anticline

112° 23' 45" 112° 20' 112° 15' 112° 10' 112° 06' 30"

39° 35' 39° 30' 39° 25'

B B' C C" G G'

TRUE NORTH 17½° MAGNETIC NORTH

Fool Creek Sinks

FOOL CREEK

LEAMINGTON

Leamington Canyon; Tank Canyon; Wood Canyon; The Pass; Yellowstone Canyon; Wild Horse Canyon; Wide Canyon; Bridge Canyon; Fork Creek; Burr Fork; Sevier River; SEVIER RANGE; SCIPIO PEAK 9763

D(?) U(?)

Undifferentiated limestone and dolomite

Anticline

112° 20'

39° 30'

39° 25'

39° 20'

TRUE NORTH

MAGNETIC NORTH

17½°

Fool Creek Sinks

FOOL CREEK

OAK CITY

Yellowstone Canyon

Wild Horse Canyon

Box Canyon

Bridge Canyon

Oak Creek

Dry Fork

Wide Canyon

Cow Canyon

John Williams Canyon

THRUST

CANYON RANGE

SCIPIO PEAK 9763

Qd

Qpb

Qa

Tl

Tf

Ki

KTn

KTn (?)

KTn (t)

pCp

Ct

Ct (t)

Cts

Cc

U D

B B'

C C'

C" G G'

D D'

E E'

Base from Soil Conservation Service Aerial Photographs, Compiled by U.S. Geological Survey

Geology by F. W. Christiansen

1 0 1 MILE

Scale in miles

GEOLOGIC MAP OF CANYON RANGE AREA, UTAH

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 63, PP. 717-740, 2 FIGS., 5 PLS. JULY 1952

STRUCTURE AND STRATIGRAPHY OF THE CANYON RANGE, CENTRAL UTAH

By F. W. Christiansen

Abstract

Stages in the development of the Canyon Range have been determined from structural and stratigraphic relationships of the following sedimentary units: Precambrian (?) quartzite, shale, and slate, 7950 feet (thrust contact at base of section); Tintic quartzite, 1500 feet (Lower Cambrian); Ophir shale, 975 feet (Middle Cambrian); undifferentiated limestones and dolomites, 4750 feet (Upper Cambrian and Ordovician?); Indianola (?) group, 12,500 (Upper Cretaceous); North Horn (?) and Flagstaff (?), 3500 feet (late Cretaceous and Paleocene); Oligocene conglomerates, 1200 feet; and Pleistocene and Recent alluvium. The following series of events is indicated.

1. Uplift accompanied by folding and large scale thrusting beginning in late Jurassic (?) time and culminating in early Cretaceous time;
2. Deep erosion, 5000–10,000 feet;
3. Deposition of thick series of coarse clastic continental sediments (Indianola group);
4. Folding and high-angle thrust faulting in post Indianola time, previous thrust plane folded and Precambrian (?) rocks thrust over Cretaceous rocks, displacement less than 5000 feet;
5. Deposition of continental sediments in marginal basins, (North Horn and Flagstaff);
6. Moderate folding in post Flagstaff time;
7. Development of essentially the present drainage systems;
8. Deposition of Oligocene (?) conglomerates on piedmont slopes and in canyons and valleys within range, 0–1800 feet thick;
9. Tilting of Oligocene conglomerates;
10. Basin and Range faulting, beginning in Miocene (?) time and continuing to Recent;
11. Alluviation of piedmonts and deposition of Bonneville sediments followed by post-Bonneville dissection.

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Introduction

Location of Area

The Canyon Range is located in the west central part of Utah about 90 miles south of Salt Lake City (Fig. 1). It lies at the eastern margin of the Great Basin province, west of the Gunnison Plateau and north and slightly west of the Pavant Range. Scipio, a small farming community, is located in a nearly circular-shaped valley southeast of the range and Oak City, Fool Creek, and Leamington are at its west base. The Canyon Range can be reached from the northeast by bus or passenger car on U. S. Highway 91, which passes through Scipio. U. S. Highway 132 and 148 from Nephi to Leamington skirt the northern extension of the range. The Tintic Mining District and West Tintic Mountains are located approximately 20 miles north of the area. The Canyon Range is a prominent physiographic feature of the area. Several of its peaks approach 10,000 feet in elevation and rise approximately a mile above the surrounding valleys. The general outline of the range is wedge-shaped, narrowing from approximately 12 miles in the southern part to approximately 3 miles in the vicinity east of Leamington. The eastern margin of the range trends approximately north-south, and the western margin north 25° east.

Purpose of Study

The principal purposes of he study were to describe the structures of the Canyon Range, determine the stages in its tectonic development, and correlate the local structures and tectonic events with those of the High Plateaus of Utah situated east of the range and with those of the western part of the Great Basin province.

Previous Literature

The Canyon Range has hitherto received no detailed geologic study. Gilbert (1890, p. 164–166, 192–193) investigated the features produced by Lake Bonneville east and west of the range and in Leamington Canyon, but paid no particular attention in his report to the geology of the range, except to indicate in a structure section and in his text that the Lake Bonneville sediments "abut against the steep face of the quartzite constituting the mountain front." He lists the bedrock of the range as "Paleozoic Sandstone."

Tower and Smith (1899, p. 617, 671, 673), reporting on the geology and mining industry of the Tintic District, referred to folded Paleozoic strata including "Carboniferous" limestone exposed in Leamington Canyon and to an unconformity between a conglomerate, which they considered of Eocene age, and folded Paleozoic rocks. Davis (1905–1908, p. 28–34) published a report on the Canyon Range after a short excursion to the area. His trip took him through Leamington Canyon, thence southwest along the western margin of the range to Oak City. From Oak City he traveled eastward to the head of Oak Creek and to the crest of the range. His sketch structure sections of Oak Creek and Dry Creek show folded limestones and quartzites which, he states, "are believed to be of Carvoniferous age." Loughlin (1912, p. 448) visited Leamington Canyon later, and observed the unconformity between the conglomerate, which he considered of Eocene age, and Paleozoic strata. He noted that the quartzites and limestones cropping out in the canyon closely resembled the lithologic character of Cambrian strata in the Tintic District and therefore concluded that they were probably Cambrian.

Loughlin (1914, p. 51–66) made a brief reconnaissance study of the area and published a

report that included a reconnaissance map of the northern part of the range. His map shows the limestones and quartzites to be Carboniferous and the conglomerates Eocene. His structural sections show folded Carboniferous strata below essentially horizontal Eocene strata. The present study concludes that the "Carboniferous" strata are Precambrian and Cambrian, and that the "Eocene" conglomerates are Cretaceous.

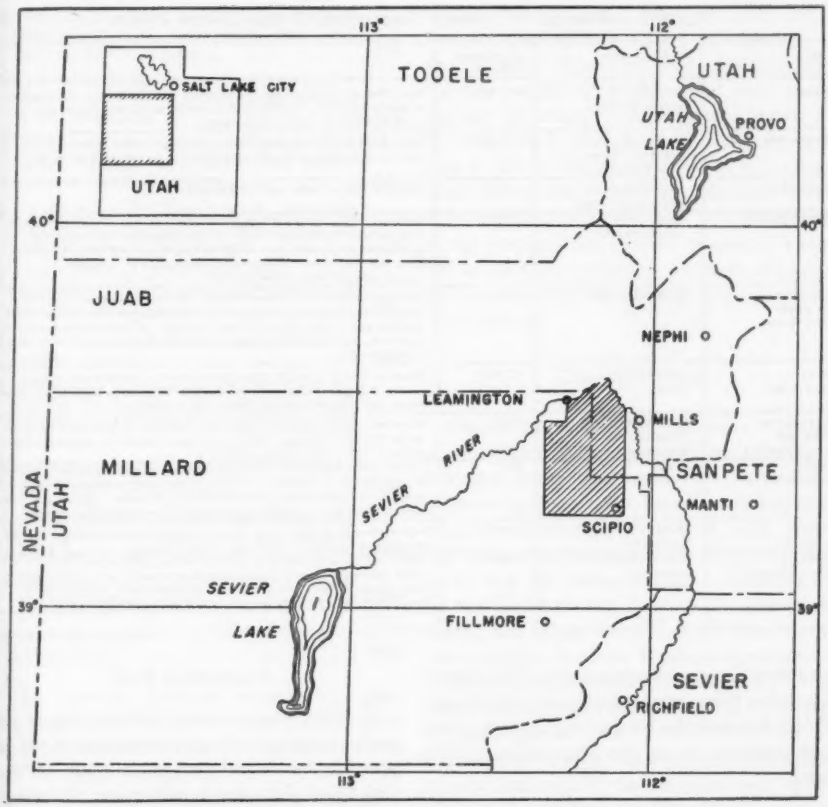


FIGURE 1. INDEX MAP TO CANYON RANGE AREA, WEST CENTRAL UTAH

Acknowledgments

This study was initiated in the latter part of the summer of 1941, under a grant from the Department of Geology of Princeton University, which defrayed the cost of aerial photographs and expenses incurred during a 6-week field period. Field study was resumed in the spring of 1946, partly financed through the United States Geological Survey-University of Utah cooperative program. The base map was prepared by the United States Geological Survey from aerial photographs obtained from the Soil Conservation Service.

The writer wishes to thank Professor W. T. Thorn, Jr., for many helpful suggestions during the progress of the field work and the preparation of the manuscript. Professor B. F. Howell identified several collections of Cambrian fossils, Professor A. J. Eardley read the manuscript and has given many helpful criticisms. Mr.

Chas. B. Hunt helped prepare the base map and Keith Erickson drafted the cross-sections and maps. For this assistance the writer wishes to express his sincere appreciation. The writer also wishes to express his gratitude to Professor Hyrum Schnieder for his untiring assistance and encouragement during the preparation of this paper.

TABLE 1.—SUMMARY OF EXPOSED FORMATIONS IN THE CANYON RANGE

Age	Formation	Thickness (feet)	Lithology
Recent	Alluvial and eolian deposits	25-500'	Post Bonneville alluvium, very small alluvial fans at the mouths of small canyons, and below Bonneville terraces; eolian quartz-sand deposits on western piedmont.
Pleistocene	Valley alluvium and Lake Bonneville deposits	25-1000'	Lake Bonneville silts and clays with interbedded alluvium; extensive pre-Bonneville alluvial deposits on margins of range.
Pliocene (?)	Lava flows and pyroclastic rocks	0-50'	Rhyolite flows, tuffs and tuff breccias.
Oligocene (?)	Fool Creek conglomerate	1800'	Light-colored, commonly mottled pale yellow, orange pink, and pale purple; composed chiefly of limestone and limestone conglomerate fragments that range in size from one inch to several feet; not well sorted, fragments subangular to rounded in matrix of calcareous sand and silt.
Paleocene and Upper Cretaceous	North Horn (?) formation	3500'	Interbedded gray to yellowish-gray medium to coarse-grained sandstone; light-gray arenaceous limestone; red calcareous shale, mottled pink, pale yellow and pale purple; calcareous and arenaceous siltstone, and gray to light-gray conglomerate; base of section red conglomerate.
Upper Cretaceous	Indianola (?) group	12,500'	Thick series of gray to red, fine to coarse, continental conglomerates that grade upward into interbedded gray to light-brown sandstone and shale with a few lentils of limestone.
Upper Cambrian and Ordovician (?)	Undifferentiated dolomite and limestone	4750'	Massive to thin bedded light-gray to dark-gray, marine limestone and dolomite; lithology similar to Upper Cambrian and Ordovician rocks of the Tintic Mining District.
Middle Cambrian	Ophir formation	975'	Lower part olive-green to light-brown micaceous shale with worm-like markings on bedding planes; 30' dark-red to black, glauconitic, persistant ridge-making quartzite at top of shale. Upper part interbedded shale and light-brown to light-gray mottled limestone.
Lower Cambrian	Tintic quartzite	1500'	Light-colored, pink to light gray, evenly bedded and thinly bedded micaceous quartzite; lower part conglomeritic.
Upper Precambrian (?)	Proterozoic rocks (?)	7955'	Thick series of alternating units of red to light olive-green shale and massive red to pale purple quartzite and conglomerate; light gray, oölitic, and stylolitic limestone beds in lower part; basal beds of section in thrust contact with younger rocks.
	Total	36,640'	

STRATIGRAPHY

General Statement

Except for a small exposure of lava flows and pyroclastic materials in the northeast part of the area, no igneous rocks have been found. The exposed sedimentary rocks are predominantly clastic with subordinate amounts of marine limestone and dolomite and lacustrine limestone. The preponderance of unfossiliferous continental and marine clastic sediments, in addition to thick sequences of unfossiliferous, fine-grained, marine dolomite and limestone, has made dating the formations and correlation difficult. However, tentative age determinations and correlations have been made for the most part, on evidence other than biologic (Table 1).

Precambrian Rocks

A thick sequence of interbedded quartzites, shales, conglomerates and limestones is well exposed in the central part of the range north of Oak Creek. This sequence is separated from the overlying Cambrian strata by a zone of relatively coarse conglomerates which were included in and mapped with the Tintic quartzite. The lowermost beds of the Precambrian are in thrust contact with younger rocks. A section measures along the crest of the ridge north of Oak Creek is as follows:

Unit		Thickness Feet
15.	Quartzite, light gray to tan, evenly bedded, medium- to coarse-grained, conglomerate lenses common	1950
14.	Shale, maroon to tan, micaceous, varies considerably in thickness	480
13.	Quartzite and conglomerate, light purple to red; quartzite coarse-grained and interbedded with conglomerate; conglomerate pebbles rounded to subangular, partley composed of vein and pegmatite quartz	1200

Unit	Thickness Feet
12. Shale, maroon, micaceous, varies considerably in thickness	75
11. Shale, tan to light brown, micaceous	300
10. Shale, light brown to gray, arenaceous in upper part with interbedded thin fine-grained quartzite beds	175
9. Quartzite, light brown to gray, evenly bedded, beds up to 2 feet thick, medium to coarse grained	1350
8. Shale and quartzite interbedded, light brown to light gray, calcareous, mostly covered, weathers to slope	335
7. Limestone, light bluish gray, massively bedded, stylolytic, some beds oölitic, locally prominent ridge-maker	60
6. Quartzite and shale interbedded, greenish-brown to tan; quartzite thin-bedded and very fine-grained	200
5. Limestone, light bluish gray, massively bedded, styolitic	140
4. Limestone, arenaceous and argillaceous, brown to tan, quartz sand laminations common, weathers to rough sponge-like surface	210
3. Quartzite, brown to greenish brown, very fine-grained, thin-bedded	400
2. Shale with interbedded quartzite, light brown to tan, shale micaceous, quartzite very fine-grained	640
1. Quartzite, light gray to tan, medium-grained, evenly bedded	440
Thrust fault contact	
Total thickness	7995

The presence of several massive light-gray limestone beds in the lower part of the section, the persistence of the relatively thin shale units, and the great thicknesses, purity and persistence of the quartzite and conglomerate units suggest that these rocks are mostly marine in origin. The red shales and perhaps part of the other argillaceous sediments may have accumulated under deltaic or floodplain conditions.

Eardley and Hatch (1940, p. 833–34) proposed a "multiple environment theory" to explain lithological variations such as tillites, altered volcanic rocks, and thinly bedded or varved sediments that occur in the sequences in northern Utah but are not present in the Canyon Range. Under this theory, the Canyon Range sequence developed in the southern extension of the Sheeprock-Cottonwood trough, but was apparently deposited in the trough sufficiently far from glaciated highlands and areas of volcanism that the accumulating sediments were not influenced.

Thus far, the local Precambrian (?) sediments have yielded no fossils, consequently their precise age has not been determined. They are, in general, lithologically similar to sequences in the Wasatch Mountains and adjacent areas. Eardley and Hatch (1940, p. 840), after studying and describing these sequences, concluded that the beds in the upper part of the sequences may be lower Cambrian and those in the lower part Upper Proterozoic. The Canyon Range sequence is structurally and stratigraphically in the same position relative to rocks of known age as are these sequences; therefore, the upper part may be Cambrian, and the lower part latest Precambrian.

Cambrian System

General statement.—The Cambrian sequences of the Canyon Range, in general similar to Cambrian rocks in adjacent areas (Table 2), are of three lighologic types: lower quartzites, intermediate limestones interbedded with shales, and upper limestones and bolomites. Lithologically, they are similar to rocks of the same age in the Tintic Mining District 20 miles to the north, the House Mountains 45 miles to the west, and the southern Wasatch Mountains 25 miles to the northeast. Good exposures in the Canyon Range can be seen along the north slope of Oak Creek.

Loughlin (1914, p. 24) made a reconnaissance survey and map of the northern part of the Canyon Range. He mapped the Tintic quartzite and overlying limestones of Cambrian age as Carboniferous. In the Yellowstone Canyon area, where he made his closest observations, the section is overturned to the east and is now dipping at a low angle westward. In regard to this overturned section of Cambrian rocks, Loughlin states: "No fossils were found in the quartzite, but its apparent conformable position above the limestone of Madison age suggests that its lower part at least is Mississippian."

Tintic quartzite.—The Tintic quartzite is a light-gray to pink, evenly and thinly bedded, fine-to medium-grained quartzite. It is composed of well sorted sub-angular to rounded grains of silica and is unusually pure. Thin con-

glomerate beds and lenses are common in all parts of the formation. The pebbles of the conglomerates are composed of white vein and pegmatite quartz with a few pink to light-brown

conformably below the Ophir formation from which fossils of Middle Cambrian age were collected, suggests that it developed during Early Cambrian time.

TABLE 2.—CORRELATION OF CAMBRIAN SECTIONS

Geologic Age	House Range[1]	Canyon Range[2]	Tintic District[3]	Cottonwood Area[4]	Blacksmith Fork[5]
Ordovician		(?)	Ajax limestone		
			unconformity	unconformity	
Upper Cambrian	9. Weeks limestone 1940'	1. Cambrian and undifferentiated limestones and dolomites 4750'	8. Opex dolomite 390'	3. Maxfield limestone 570'	St. Charles formation 400'
Middle Cambrian	8. Marjum limestone 1530' 7. Wheeler shale 350' 6. Swasey limestone 395' 5. Dome limestone 310' 4. Howell limestone 835'	Bonnia fauna 2. Ophir formation 975'	7. Cole Canyon dolomite 510' 6. Bluebird dolomite 175' 5. Herkimer limestone 235' 4. Dagmar limestone 100' 3. Teutonic limestone 556' 2. Ophir formation 355'	2. Ophir shale 420'	Nounan dolomite 900' 5. Bloomington limestone 1273' 4. Blacksmith dolomite 450' 3. Ute limestone 685' 2. Langston limestone 575'
Lower Cambrian	3. Tatow limestone 165' 2. Pioche shale 265' 1. Prospect Mt. quartzite 1000'	(?) 1. Tintic quartzite 1500'	1. Tintic quartzite 6000'	1. Tintic quartzite 800'	Brigham 1000' quartzite (?)
	Base covered	Unconformity (?)	Base not exposed	unconformity	
Precambrian		Precambrian		Cambrian (?) tillite	

[1] Diess (1938, p. 1141). [2] This report. [3] Loughlin (1919, p. 30). [4] Calkins and Butler (1943, pl. 5). [5] Deiss (1938, p. 1117).

quartz pebbles. They are rounded to subangular and range up to 1½ inches in diameter. The even bedding of the Tintic quartzite and its purity and persistence indicate that it was originally, in all probability, a beach sand. The conglomerate beds and lenses at the base of the section are composed of the same kind of material or fragments as those within the formation, but are thicker and more persistent and the material is coarser. These coarser conglomerates appear to rest conformably on the older quartzites, and may not represent the basal beds of the Cambrian. They were, however, mapped with the overlying quartzites and have tentatively been considered as the basal conglomerates of the Tintic quartzite. The upper contact was arbitrarily chosen at the top of the uppermost, relatively pure quartzite beds. Conformably above these quartzite beds are argillaceous and glauconitic sandstone beds of the Ophir formation.

The Tintic quartzite in the Canyon Range is unfossiliferous, and therefore its precise age has not been determined. Its stratigraphic position,

Ophir formation.—The Ophir formation of the Canyon Range is approximately 975 feet thick and is composed predominantly of limestone. Subordinate units of tan to olive-green shale are interbedded with the limestone. Near the base of the formation is a persistent and prominent glauconitic, quartzite unit that is approximately 27 feet thick, weathers brown to black on its surface, and forms low, easily recognizable hogback ridges. This member of the Ophir formation has been named the "Black Jack" by local prospectors. Below this quartzite are olive green, arenaceous and micaceous shales interbedded with brown sandstone and greenish brown, impure, fine-grained, thin-bedded quartzite units. On bedding plane surfaces or partings, the shale has characteristic wormlike markings similar to those found in most localities where the Ophir formation has been observed and described.

The upper 715 feet of the Ophir formation is predominantly limestone with several thin units of shale interbedded. The uppermost limestones are relatively pure, light bluish gray, massively

and indistinctly bedded. The limestone units are more resistant to weathering and consequently form steep slopes and ledges and locally small hogbacks. Lower in the section, the limestones become impure and are darker. Striped, mottled, and spotted surface markings are common. The color of the markings varies through shades of gray, red, brown, and yellow. The interbedded shale is olive green to light yellowish brown, has a characteristic micaceous sheen on bedding surfaces, and contains numerous very thin layers of arenaceous and argillaceous limestone, some of which were fossiliferous. Fossils collected from the uppermost shale were reported by B. F. Howell (personal communication) to be from the lower part of the middle of the Middle Cambrian. The following is a section measured on the north slope of Yellowstone Canyon, located in the northwest part of the range:

Unit	Description	Thickness Feet
17.	Shale, gray to brownish gray, calcareous, interbedded with very thin beds of dark-gray limestone, mostly covered	144
16.	Shale, olive-green to tan and brown, micaceous, weathers to low slopes, thin arenaceous limestone beds contain fossils	56
15.	Limestone, light bluish gray, thin-bedded, light-brown to yellow shale interbedded	6
14.	Limestone, light bluish gray on weathered surface, pinkish gray on fresh surface, fine-grained, massively bedded, and indistinctly bedded, ridge-maker	55
13.	Limestone, gray with light-gray ribbon-like markings that parallel the bedding planes, evenly bedded, fine-grained, partly covered	162
12.	Shale and limestone interbedded; shale, olive green, micaceous and calcareous; limestone, gray, thinly bedded, fine-grained	97
11.	Limestone, light brown to tan, dense, laminated, locally contains breccia	7
10.	Limestone, gray, limonite stain along wavy bedding planes, thin to massively bedded, some beds 4′ thick	17
9.	Limestone, light gray to gray, massivly bedded, irregularly bedded, light-brown irregular-shaped markings on weathered surface	65
8.	Limestone, gray; lower 10 feet mottled with light-gray markings, weathers to scoria-like surface; central part contains dark-gray to reddish-brown concretionary-like structures ¼ to 1 inch in diameter; upper part similar to lower 10 feet	44
7.	Limestone, light brown, thinly bedded, few thin beds of olive-green shale	12
6.	Limestone, gray with brownish-gray irregular knob-like structures on weathered surfaces, evenly bedded, weathers to slope	9
5.	Limestone, gray, very dense, weathered surface irregularly banded, some beds have tan to brown patches and spots	10
4.	Limestone, gray, arenaceous in lower part grading up to pure crystalline limestone, evenly bedded, weathers to slope	39
3.	Shale and calcareous sandstone interbedded; sandstone brown to tan on weathered surfaces, greenish-brown on fresh surfaces, glauconitic; shale like unit 16	170
2.	Quartzite, brown to black on weathered surface, light green on fresh surfaces, glauconitic, evenly bedded, ridge-maker, locally known as the "Black Jack"	27
1.	Sandstone and shale interbedded, olive green to tan, micaceous, evenly bedded; arenaceous shale bedding surfaces have wormlike markings; glauconitic	55
	Total	975

The Ophir formation varies as much as 400 feet in thickness within the Canyon Range, for the most part, to the local secondary effect of compression. Some variation may be due to differences in the original thicknesses of the sediments, although this has not been demonstrated. The lower and upper boundaries of the formation are conformable with the beds below and above. The upper contact or boundary of the Ophir formation was arbitrarily drawn at the top of the upper fossiliferous shale unit.

Upper Cambrian and younger undifferentiated limestone and dolomites.—Conformably above the Ophir formation is a thick series of light- to dark-gray limestones and dolomites. In general they are dense, hard, even-bedded, unfossiliferous, marine sediments, displaying a wide variety of peculiar yet characteristic lighologic features and surface markings. These features are common to Upper Cambrian and younger rocks of adjacent areas. The upper beds in the central and southern part of the range are in thrust contact with rocks of Proterozoic (?) age. Along the northern slopes of Oak Creek Canyon the section reaches its maximum thickness. A measured section in this vicinity, and in Yellowstone Canyon, revealed a total thickness of 4750 feet.

Since no fossils have been found in these rocks, their age has not been accurately determined. However, they are strikingly similar

lithologically to limestones and dolomites that occur above the Ophir formation in the Tintic District, southern Wasatch Mountains, Pavant Range, and other localities. They occur immediately above and conformably on rocks of lower Middle Cambrian age; therefore, the lower part of the sequence is very likely equivalent to the upper Middle Cambrian and Upper Cambrian rocks of the Tintic District and other adjacent areas. The uppermost beds of the section may be Ordovician or younger. On the southeastern front of the range, a thick section of predominantly marine chemical sediments are present. The lower limestones of this section are lithologically similar, and may be equivalent to the Garden City formation of Ordovician age. Above these limestones is a thin quartzite unit (50 feet) overlain by dark-gray crystalline dolomite. The thin quartzite and dark-gray dolomite may be Ordovician and if so probably correlate with the Swan Peak formation and Fish Haven dolomite respectively. Above the dark-gray dolomite is several hundred feet of light-gray, dense, even-bedded dolomite which is lithologically similar to the Laketown dolomite of Silurian age.

The following section measured on the north side of Oak Creek Canyon, and in Yellowstone Canyon, gives some of the lithologic details and thicknesses:

Unit	Description	Thickness feet
23.	Limestone, light gray, fine-grained, slope-maker, mostly covered	380
22.	Dolomite, gray to light brown and tan, fine-grained, brown to dark-gray chert nodules and lenses common, veined with calcite, slope mostly covered	156
21.	Dolomite, gray, fine- to coarse-grained, forms steep slopes, mostly covered	282
20.	Dolomite, light to dark gray, hard, dense, fine- to coarse-grained, evenly bedded, uniformly bedded, few beds of limestone	590
19.	Limestone, gray to yellowish gray, fresh surfaces mottled brown and yellow, mostly covered	310
18.	Covered steep slope, light- to dark-gray, fine-grained dolomite fragments compose talus	140
17.	Dolomite, light to dark gray, massively bedded, straight curved and branching white calcite and dolomite rodlike structures abundant in dark dolomite, some beds contain dark-gray pisolitic-like structures enclosed in light-gray to white shells, in medium-gray matrix	120
16.	Shale, tan to brown, micaceous, calcareous, wormlike markings on bedding partings	22
15.	Dolomite, gray on fresh surfaces, weathered, massively bedded, medium-grained	75
14.	Dolomite, light-gray, vuggy irregularly shaped openings lined with calcite and dolomite crystals, bedding planes indistinct	18
13.	Dolomite, brown to tan, thinly bedded, arenaceous, limonitic	36
12.	Dolomite, dark gray, fine-grained, "spangled" with white rodlike structures, thin to massively bedded	298
11.	Dolomite, light- and dark-gray strata, interbedded dark-gray beds contain rodlike structures similar to Unit 12, medium to massively bedded	118
10.	Dolomite, light to dark gray, fine- to coarse-grained, thin to massively bedded, white rodlike structures common, some light-gray beds on weathered surface show laminated structure	772
9.	Covered, float dark-gray dolomite and limestone	152
8.	Limestone, light gray, laminated	25
7.	Limestone, gray to dark gray, evenly bedded, light- gray ribbonlike markings on weathered surface, bedding planes display wormlike markings, white calcite rodlike structures common	115
6.	Dolomite and limestone, light-gray laminated dolomite interbedded with massive-gray limestone	131
5.	Dolomite and limestone, light to dark gray, evenly bedded, upper light-gray dolomite and limestone thinly bedded and laminated	270
4.	Limestone, gray, evenly and uniformly bedded, upper 35' oölitic, red veinlets of calcareous clay and calcite parallel and cut across strata, some beds banded and mottled lighter and darker shades of gray, white rod-like structures in dark-gray beds	212
3.	Limestone, dark gray, lower few feet light gray, ribbonlike banding common, blotches and veinlets of limonite-stained argillaceous material occurs along joint planes and bedding planes, middle part covered	366
2.	Limestone, gray, massively bedded, weathered surfaces lighter gray, upper part has ribbonlike bandings	82
1.	Limestone, gray to light gray, dense fine-grained, light-gray crinkly banding, some light-gray beds laminated	79
	Total	4749

Cretaceous Rocks

General statement.—Cretaceous beds rest unconformably on Upper Cambrian strata at the head of Yellowstone Canyon. Southward from this locality, they truncate progressively older and older beds, until they come to rest on rocks of Precambrian (?) age. The greatest thickness and most available exposures of the Cretacious beds occur in the northeastern part of the range between Wide Canyon and the Pass Canyon (Pl. 2, Sec. BB). Here they reach a total thickness of approximately 12,500 feet. This thick sequence is composed predominantly of coarse clastic sediments of continental origin and has been tentatively correlated with the Indianola group of the Gunnison Plateau. The lower conglomerates of the sequence are composed of limestone and quartzite cobbles and boulders. Upward in the section, the sediments generally become less coarse, grading into pebble conglomerates, coarse to fine-grained sandstone, sandy shale, and in a few places thin beds of fresh-water limestone. From west to east there is also an apparent gradation to less coarse sediments. This gradation, in addition to the cross-bedding in the sandstone strata, indicate that the sediments were derived from a mountainous terrain west of the present outcrops.

Unconformably above these continental sediments is a finer and much thinner (3500 feet) sequence of continental strata. This sequence consists of a highly colored, heterogeneous assemblage of interbedded conglomerates, sandstones, siltstones, and impure fresh-water limestones. Tentatively, they have been determined to be of late Cretaceous (Lance) and Paleocene ages, although the lower-most strata may be of Late Montana age. Definite age determinations could not be made because of the unfossiliferous nature of the sediments. Lithologic similarity and stratigraphic position relative to rocks of known age were used in dating them.

Indianola (?) group.—Tentative correlation of the Cretaceous conglomerates with the Indianola group of the Wasatch and Gunnison plateaus east of the Canyon Range has been made on the basis of their general similarity in lithology and approximate stratigraphic position relative to rocks of known age. Immediately east of the Canyon Range, in Long Ridge, Siegfried Muessig (personal communication) has identified the North Horn, Flagstaff, and Green River formations. The lower part of this sequence, the North Horn formation, and probably part of the Flagstaff formation, crop out in the eastern foothills of the Canyon Range in the vicinity of Mills Road, and is there resting unconformably above the thick sedimentary sequence here correlated with the Indianola group. The basal beds of the Indianola (?) group of the Canyon Range are unconformable on folded Cambrian and older strata. The surface over which the conglomerates were deposited was one of moderate relief. In the vicinity of Scipio Peak, the basal conglomerate beds were deposited in strike valleys that developed in pre-Indianola time on the less resistant shale and conglomerate beds. Exposed contacts reveal a relief of approximately 800 feet.

Several coral specimens of Mississippian age were found in the pebbles and cobbles of the lower limestone conglomerates, and fragments of sandstone were noted that are lithologically similar to the Navajo sandstone. The evidence at hand indicates that these sediments here called the Indianola (?) group, were deposited prior to the deposition of the North Horn formation and after the deposition of the Navajo formation. It is probable, therefore, that the basal beds of this sequence may be of latest Jurassic (Morrison) age.

Immediately north of Wide Canyon and at the head of Yellowstone Canyon, the Indianola (?) sediments reach a maximum development and are well exposed. In this vicinity they can be broken down into three fairly distinctive units: a lower part consisting predominantly of Paleozoic limestone fragments, a middle part consisting of massively bedded quartzite cobble and boulder conglomerates, and an upper part of less coarse conglomerates interbedded with sandstone, siltstone, shale, and a few fresh-water limestone beds and lenses.

At the head of Yellowstone Canyon the lower unit is 800 feet thick and consists predominantly of limestone fragments enclosed in a calcareous, sandy to pebbly matrix. The lower beds are less coarse and are intercalated with beds and lenses of red calcareous sandstone. Approximately 90 per cent of the fragments were derived from Paleozoic limestones that are readily recognized by their peculiar lithology. Cam-

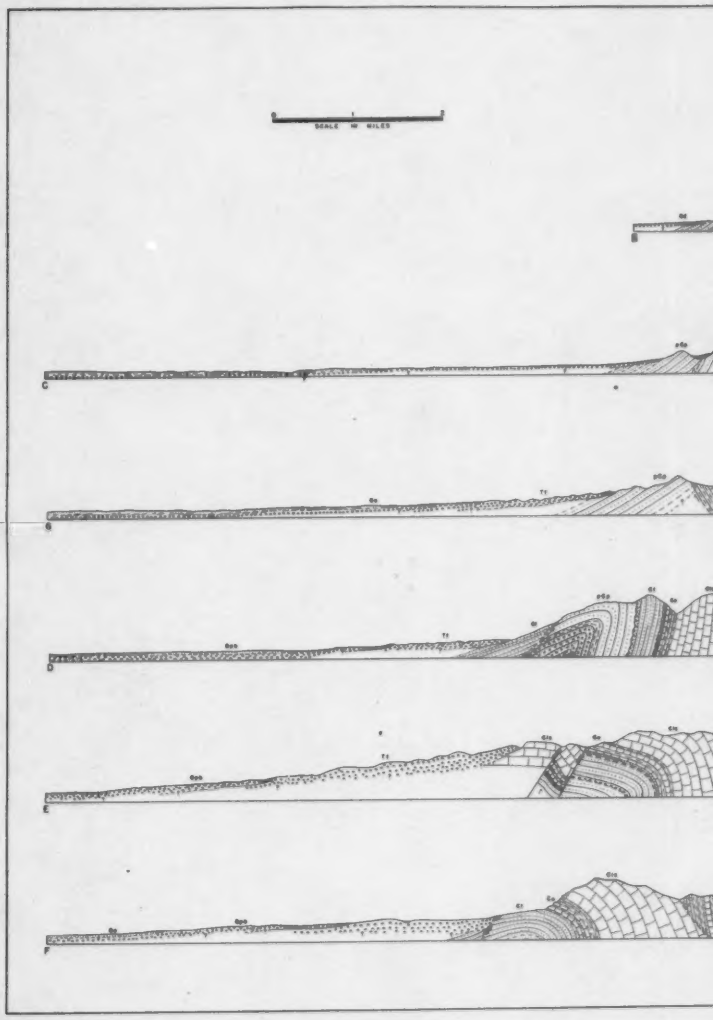
brian lithologic types predominate. However, there are subordinate amounts of Carboniferous and unrecognizable fragments. Of the remaining 10 per cent, most are fragments that resemble Tintic and Precambrian (?) quartzites. A few brown to red and light-gray to white fragments of sandstone were found that are lithologically similar to Triassic and Jurassic rocks exposed in areas south of the Canyon Range. The brown to red sandstone is similar to the sandstone occurring in the upper sandstone strata of the Chinle formation and the light-gray to white sandstone closely resembles Navajo lithology. The sand grains of the light-colored sandstone are moderately well rounded, well sorted, and are frosted. It is probable that the Navajo sandstone was involved in the pre-Indianola folding and thrusting, and consequently supplied part of the material that was shed from the folded and faulted mountainous terrain that existed in and west of the Canyon Range area.

Above this limestone conglomerate is a series of massive, quartzite conglomerate beds, approximately 1800 feet thick, that make up the middle unit. They are very resistant to erosion and consequently develop precipitous ledges and steep slopes. The lower beds contain less coarse fragments than the central and upper parts and are devoid of limestone pebbles or cobbles. Higher in the section limestone fragments increase in amount up to 15 per cent near the top. The central part is an extremely coarse conglomerate that contains boulders up to 7 feet in diameter. The boulders are composed of red to purplish-red, pink and light-gray quartzzite derived from the Tintic quartzite and Proterozoic (?) quartzites. They are angular, subangular and rounded and are embedded in a matrix of pebble conglomerates, coarse to fine sandstone, and siltstone.

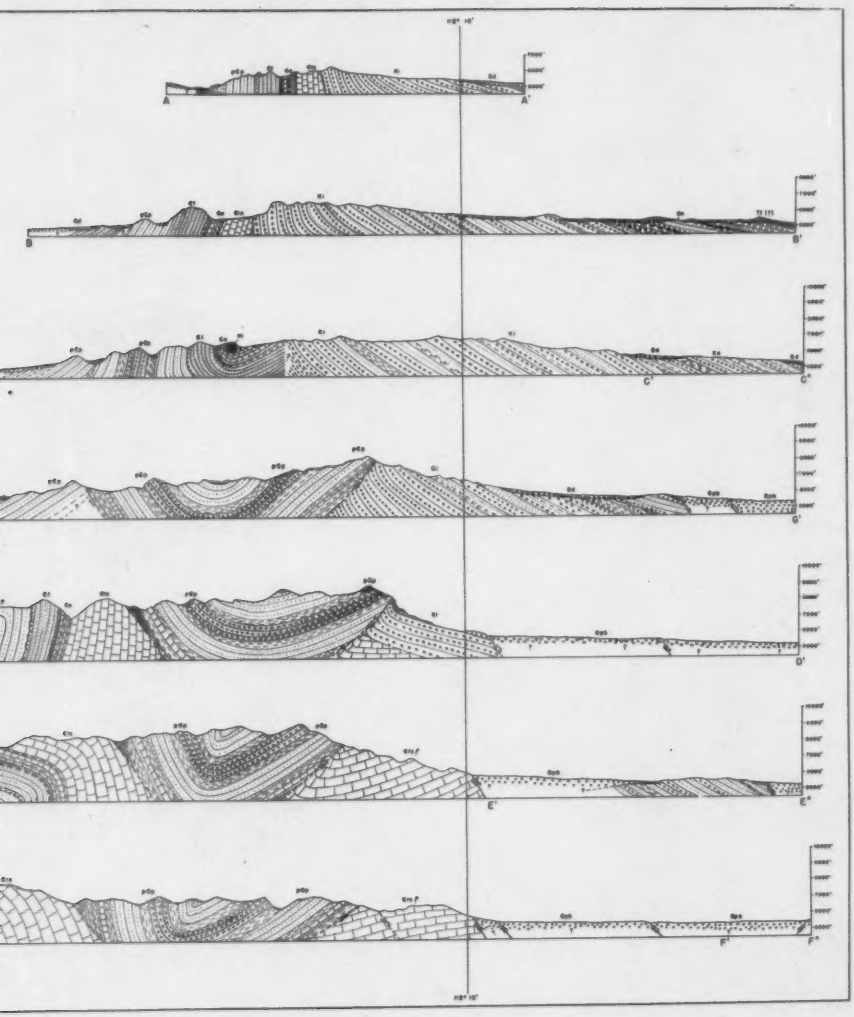
The upper unit reaches a maximum thickness of 9000 feet in the area north of Wide Canyon. Relative to the underlying strata, these beds contain clastic materials that are considerably more sorted and rounded and much smaller in size. Interbedded conglomerates, sandstones, siltstones, and shales make up the greater part of the unit. Near the upper part are several beds of freshwater lithographic-like tan to brown limestone. The conglomerate strata range from nearly pure limestone conglomerate through varying mixtures of limestone and quartzite fragments to nearly pure quartzite conglomerate. The sandstone beds are gray to yellowish brown and the sand is poorly sorted and generally argillaceous. The shale and siltstone beds are generally calcareous and vary greatly in color. The beds high in carbonate are usually light gray to white; less calcareous beds are tan to brown and red. A few sandy and calcareous siltstones near the top of the section are mottled pink, light yellow and shades of purple.

Specific correlation of these units of the Indianola (?) group with other exposed sequences in adjacent parts of the Canyon Range is not possible because of the discontinuous or lenticular nature of the strata. The lower 800-foot limestone conglomerate unit, for example, lenses out completely within a distance of 4 miles. Individual beds, as would be expected, lense out in shorter distances. This characteristic, among others of the Indianola (?) group, points to an environment of deposition one would expect to find along the margins of a relatively large and vigorously uplifted mountainous highland.

Beginning in late Jurassic (?) time and extending into the early Cretaceous, the older rocks of the region were folded, thrust-faulted, and uplifted. The highlands produced by this orogeny shed enormous quantities of debris which accumulated along the margins of the uplift and in adjacent intermontane basins. The present site of the Canyon Range was located at the eastern margin of one of perhaps several uplifted mountainous masses that developed in central and western Utah and eastern Nevada in pre-Indianola time. Streams flowing eastward from a highland lying west of the Canyon Range area carried large volumes of debris which was partly deposited in the form of alluvial fans along the piedmonts of the highland. The finer material was carried to the lowland beyond, where it accumulated in stream channels and on flood plains. Within the floodplain belt and also adjacent to the source of sediments, lakes appeared in which silt, clay, and limestone were deposited. Some lakes probably originated by the adjustment of stream patterns resulting from the rapid accumulation of sediments. Lakes that appeared near to the source of sediments were probably due to the coalescing of alluvial fans, as suggested by



GEOLOGIC CROSS SECT



CROSS SECTIONS OF CANYON RANGE

Eardley (1932, p. 406) for the origin of similar limestone deposits in the Wasatch conglomerate.

During the progress of erosion of the highland and filling of the adjacent basin, the expanding area of sedimentation progressively overlapped onto the deeply eroded borders of the highland. This process continued until the present area of the Canyon Range was buried beneath a thick wedge of Cretaceous and probable late Jurassic sediments (Pl. 3).

Cretaceous and Tertiary Systems

North Horn (?) formation.—Unconformably above the Indianola (?) group occurs a heterogeneous assemblage of strata which is generally lighter and more varicolored and much finer than the underlying sediments. This series of strata consists of interbedded arenaceous limestones calcareous sandstones, siltstones, and conglomerates that aggregate 3500 feet in thickness. They are tentatively correlated with the North Horn formation of the Gunnison Plateau. Spieker (1946, p. 132) named and described this formation as follows:

"The name is derived from the type locality, on North Horn Mountain . . . (on the eastern slope of the Wasatch Plateau). In the western part of the plateau, the formation consists of the former lower Wasatch, exclusive of the basal conglomerate, which now is placed in the Price River formation. The North Horn includes an assortment of lithologic types hitherto considered regionally characteristic of the Tertiary—variegated shale and sandstone, conglomerate, freshwater limestone and other beds. . . ."

The North Horn (?) formation of the Canyon Range area is partly exposed in the piedmont belt northeast of Wide Canyon. Here the upper strata of the sequence disappear eastward beneath Pleistocene and Recent sediments of the Sevier River valley. Inmediately east of the Sevier River, at the base of Long Ridge (not included in the area mapped), the upper strata of the North Horn formation are exposed and there occur stratigraphically below the Flagstaff formation of late Paleocene age. Precise correlation of the North Horn strata of Long Ridge with the North Horn (?) formation of the Canyon Range, several miles to the west, has not been made. However, their general lithologic similarity suggests that they may be contemporaneous. This is particularly true of the strata in the central part of the North Horn (?) formation of the Canyon Range. It is possible, however, that the upper beds of this sequence may be equivalent in age to the lower portion of the Flagstaff formation, and the lower part may correlate with the upper part of the Price River formation. In the absence of diagnostic fossils or other more definite evidence, this cannot be verified.

The predominant type of limestone of the North Horn (?) formation is gray to nearly white, arenaceous, relatively dense, and occurs in beds and lenses that are several feet thick. A few light-tan to brown argillaceous and arenaceous beds are present in the upper part of the section. Shale and siltstone comprise 50 per cent of the formation. The shale beds are dark red to light red and tan, and usually break down into dark-red soils. The siltstones and calcareous siltstones vary considerably in color. They are commonly light brownish red, light gray to pinkish, or light shades of lavender. The arenaceous and calcareous siltstones are usually mottled in pastel shades of lavender, red, and yellow. The interbedded sandstone is chiefly light gray to tan and yellowish-brown, calcareous, medium- to coarse-grained and unsorted. The grains in the sandstone are angular to subangular. The thickest beds and lenses usually contain iron concretionary masses unevenly distributed through the sandstone. Conglomerate lenses occur commonly within the sandstone strata. Conglomerate units are generally less than 15 feet thick, although a few may be 30 feet thick. They are made up of subangular to rounded pebbles and cobbles. The composition of the individual conglomerate beds varies from essentially pure limestone conglomerates through quartzite and limestone conglomerates to pure quartzite conglomerates.

Oligocene (?) Series

Fool Creek conglomerate.—The Fool Creek conglomerate is a new formational name and is here defined to include conglomerate beds exposed in many parts of the range, but it is particularly well developed in Dry Fork of Fool Creek and in a long, gentle spur projecting westward from the Canyon Range in the area

immediately south of the small farming community of Fool Creek. It is composed of pebbles, cobbles, and boulders derived from most of the older formations exposed in the range and in the piedmont adjacent to the range. The fragments are angular to subangular, and range up to 4 feet in diameter. The matrix is mud, calcareous sand, grit, and pebble conglomerate. The formation is generally well lithified, but bedrock outcrops are not common and occur only where erosion is relatively vigorous, such as at the heads of gullies or in areas of recent cloudburst floods.

Characteristically, the Fool Creek conglomerate erodes to smooth, rolling hills, such as are present in Dry Fork of Fool Creek and in the foothills south of the town of Fool Creek. Although it at one time covered a large part of the range, filling the canyons and extending up onto the summit areas, outcrops are now confined to patchlike exposures on the foothills, in the canyons, and on the broad divides along the crest of the range (Pl. 1).

The physiographic setting at the time the gravels which now compose the Fool Creek conglomerate were laid down had a profound influence on the ultimate thickness, and particularly the variation in thickness, of this formation. The evidence indicates that the gravels of the conglomerate filled the canyons and valleys to depths exceeding 800 feet, and accumulated along the margins of the range in the form of alluvium to thicknesses exceeding 1800 feet; thus, the lenticular nature of the deposits and their extreme variation in thickness and composition.

The coarse clastic nature of the Fool Creek conglomerate, the absence of well sorted materials, its indistinct bedding, its mud, sand, and pebble conglomerate matrix, the angular to subangular nature of the fragments, and its occurrence on an irregular surface suggest that it was deposited very rapidly by fluctuating streams mainly in the form of alluvium and mudflows.

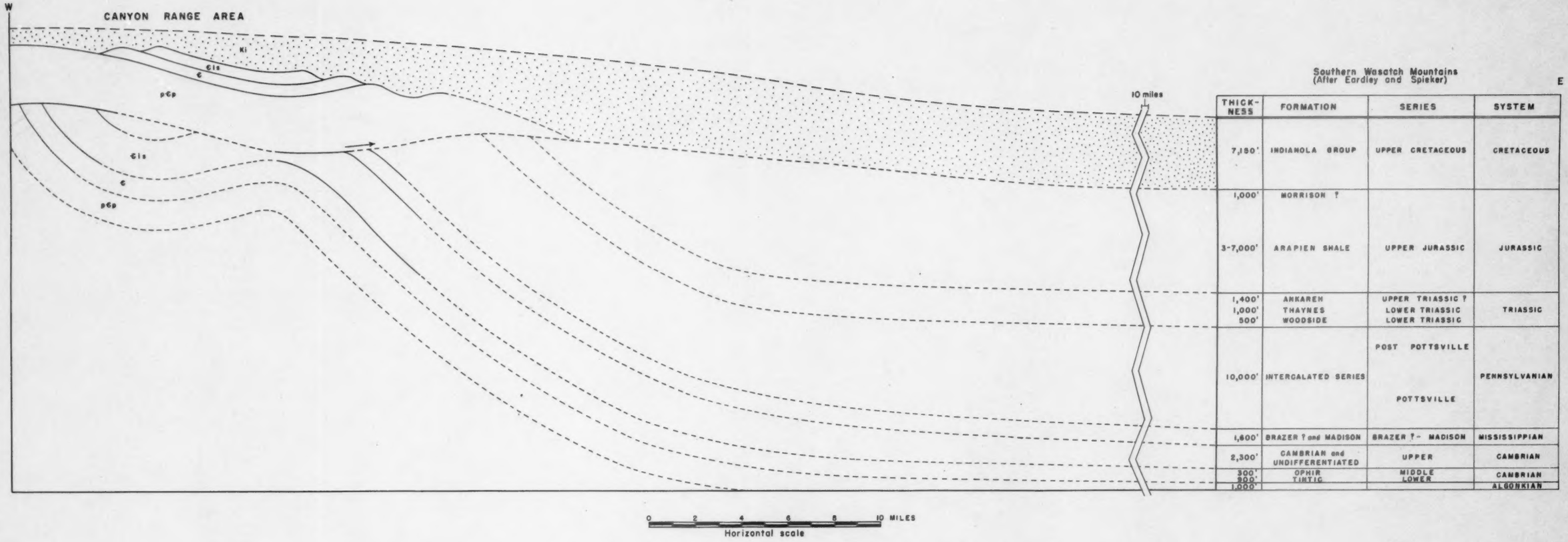
Since no fossils have been found in the Fool Creek conglomerate, it is impossible to make a definite age assignment, but its physiographic setting, lithology, and structural and stratigraphic relationship to associated rocks help to narrow the time range. It rests unconformably on the North Horn (?) formation of latest Cretaceous and earliest Tertiary ages, and has been displaced by Basin and Range faulting, the age of which has not been fixed in the area of the Canyon Range. Various geologists have placed the time of the initiation of Basin and Range faulting in the eastern Great Basin of Utah from middle Miocene to post Pliocene time. It is probable that Basin and Range faulting did not begin over the entire Great Basin Province at the same time, consequently the divergence of dates and the necessity, if faulting is to be used in dating the conglomerate, of determining the age of initiation of faulting within the area concerned.

Eardley (1933, p. 243) advanced a theory that a late youth to early mature surface with relief up to 3000 feet existed in the Wasatch Mountain area before Basin and Range faulting began. This surface is present in several localities in the Canyon Range and has developed on the Fool Creek conglomerate. The conglomerate therefore is considerably older than the pre-fault surface, which in the north-central Wasatch probably developed during Miocene (?) time (Eardley, 1944, p. 877). The age of the Fool Creek conglomerate, therefore, lies between late Paleocene and early Miocene time, and since a relatively long period of erosion occurred subsequent to the deposition of the North Horn formation and preceding the deposition of the conglomerate, it appears most likely that the Fool Creek conglomerate was deposited during Oligocene time.

Pleistocene and Recent Deposits

Pleistocene and Recent sediments occur at the margins of the range and in recent stream channels. They consist of unsorted mud flow material, alluvial gravel, sand and silt, river sand and silt, lacustrine clay and silt, and eolian sand.

Pleistocene sediments can be divided into two types: pre-Bonneville alluvial fan and mud flow deposits, and Lake Bonneville clays and silts. The western and eastern piedmonts of the range are composed primarily of pre-Bonneville deposits that have been terraced by wave action and along shore currents of Lake Bonneville. These deposits are crudely stratified and are



THICKNESS	FORMATION	SERIES	SYSTEM
7,150'	INDIANOLA GROUP	UPPER CRETACEOUS	CRETACEOUS
1,000'	MORRISON ?		
3-7,000'	ARAPIEN SHALE	UPPER JURASSIC	JURASSIC
1,400'	ANKAREH	UPPER TRIASSIC ?	
1,000'	THAYNES	LOWER TRIASSIC	TRIASSIC
500'	WOODSIDE	LOWER TRIASSIC	
10,000'	INTERCALATED SERIES	POST POTTSVILLE POTTSVILLE	PENNSYLVANIAN
1,600'	BRAZER ? and MADISON	BRAZER ? - MADISON	MISSISSIPPIAN
2,300'	CAMBRIAN and UNDIFFERENTIATED	UPPER	CAMBRIAN
300'	OPHIR	MIDDLE	CAMBRIAN
900'	TINTIC	LOWER	
1,000'			ALGONKIAN

EAST–WEST SECTION SHOWING STRUCTURE OF THE CANYON RANGE AREA AT THE CLOSE OF INDIANOLA (?) TIME

made up of a variable assemblage of sediments. Pebble and cobble beds with a sandy to silty matrix predominate. The fragments are rounded to angular, and in composition correlate very closely with rock types exposed in the canyons above. Pre-Bonneville alluvium has also accumulated to a thickness of 125 feet in the large canyons that drain the western part of the range and in the piedmont belts and adjacent valleys may exceed 1000 feet in thickness.

Lake Bonneville sediments are well exposed near Leamington and in Leamington Canyon, where small, conspicuous patches or remnants are preserved. The beds are essentially horizontal and consist of arenaceous and calcareous silt and clay with interbedded tongues of alluvial gravel that thin basinward.

Alluvial fans, mud flows, and eolian sand deposits are developing on the piedmonts and in the stream channels of the range. The recent alluvial fans are confined to areas at the mouths of small canyons having steep gradients, to the downfaulted blocks of recent piedmont faults, and to localities where the intermittent streams cut through Bonneville terraces. These post-Bonneville alluvial fans are surprisingly small, and make up a very small percentage of the alluvium that has been shed from the range. West of Oak City is a field of Recent migrating and stationary dunes. They are roughly crescent-shaped and their horns point northeast. The quartz sand in them is fine-grained, and the individual sand grains are subangular to rounded. The grains are stained light red with iron oxide, which gives the dunes a dull, rusty appearance. The quartz sand was apparently derived from playa deposits farther to the west and from intermittent stream channels that cross the piedmont.

Structure

General Setting

The Canyon Range (Pls. 1, 2) lies immediately west of the transition zone between the High Plateaus of Utah and the Basin and Range province, and southwest of the central Rocky Mountains. It is of considerable geologic interest, since its tectonic history is closely related to these three geologic provinces.

The youngest structural elements of the Canyon Range (Basin and Range normal faults) have been traced continuously from the foothills of the Canyon Range into the High Plateau province of Utah. The zone of normal faults (Pl. 5) in the eastern piedmont of the range trends southward to a point where it joins the normal fault zone at the northwestern margin of the Sevier Valley graben. The folds of the range have the same general trend as the folds of the Tintic Mining District to the north and the Pavant Mountains to the south. The Canyon Range thrust fault is apparently related to the thrusts of the Sheeprock Mountains, the southern Wasatch Mountains, and the Pavant Mountains. In general, however, the major diastrophic events recorded in the rocks of the Canyon Range correlate more closely with those of the High Plateau province to the east (Pls. 1, 2).

Folds

South of Wild Horse Canyon, the core of the range is a synclinally folded klippe composed of rocks of Upper Precambrian (?) and Cambrian ages. It has been overthrust eastward onto Cambrian and Upper Cretaceous rocks during two stages of thrusting. The plunge of the axis of the syncline varies in degree and direction. North of Scipio Peak the syncline plunges approximately 4° north. South of Scipio Peak there is a local reversal in direction of plunge that persists for about 1½ miles. South of the Oak Creek area, the axis of the fold is approximately horizontal.

The strata in the western limb of the syncline, south of Whiskey Creek, are generally vertical to slightly overturned, particularly in the upper part of the fold, and the strata of the east limb in this vicinity dip approximately 35° west.

Between Whiskey Creek and Fool Creek, the strata were less intensely folded. Near the head of Oak Creek, the eastern limb of the syncline dips westward at a low angle (8°). The western limb dips eastward at angles varying from 40° to 60°. The broad, open nature of the lowermost beds exposed in the axial zone of the fold can be observed on the south slope of Oak Creek Canyon opposite the government pasture at the eastern end of the Oak Creek road. Higher in the fold, the strata are more sharply folded.

In the northern part of the range north of Fool Creek, folding was most intense. The western limb is sharply overturned to the east and has been segmented by numerous, relatively small tear faults and tear fault zones that have exerted considerable control on the location of the small canyons in this part of the range. The varying degree of rotational movement along these tear faults accounts for the abrupt change in intensity of overturning between contiguous segments of the overturned limb. The eastern limb of the syncline in this locality was largely removed by pre-Indianola (?) erosion, and the roots that remain are buried beneath Indianola (?) sediments (Pl. 2, Sec. B-B′).

North of Wild Horse Canyon, the eastern part of the range is essentially a homoclinal structure composed of rocks of Upper Cretaceous and early Tertiary ages. The eastern part of this structure is comparatively gentle, but increases in intensity westward to a point near the center of the range, where the strata are vertical to slightly overturned.

North of the Pass Canyon, the homocline and the underlying older strata of the west limb of the syncline change their general north-south trend and veer eastward with increasing intensity to Leamington Canyon, where the general trend of the structures of the range is approximately N 35° E. This general northeasterly trend continues for a considerable distance beyond the area of the Canyon Range and can be traced to the Tertiary lava field to the northeast where the strata disappear beneath the lava flows.

The anticlinal fold in the southwest part of the range is composed of Precambrian (?) and Cambrian rocks. It is locally overturned to the east and generally plunges southward. At the mouth of Oak Creek it plunges approximately 7° southward. Farther south, in the Whiskey Creek area, an abrupt reversal in the direction of plunge of the fold persists for approximately 2 miles, beyond which it resumes its southward plunge.

The locally inverted western limb of the anticline is well exposed on the slopes of the lower part of Oak Creek in the vicinity of the Forest Service camp ground. Here, the overturned limb dips 75° west and is exposed for several thousand feet. South of Oak Creek the anticline is less intensely folded. The axial trace of this anticline and the major synclinal structure of the core of the range are approximately parallel but locally deviate as much as 12°.

Thrust Faults

The Canyon Range thrust fault is excellently exposed on the eastern slope of the range between Dry Fork and the south end of the range, (Pl. 4) and in the central and south central part of the range between the head of Bridge Canyon and the south end of the range. The regional trend of the east and west segments of the fault is generally north-south. The many small deviations in trend of the thrust shown on the map are due chiefly to the influence of topography. The west segment of the fault in the central part of the range dips east at an average of about 45°, and the eastern segment dips west at about the same angle. There are, however, some local and small variations in the angle of dip of the thrust surface. One of the most pronounced changes in dip occurs in the vicinity of Cow Canyon. The thrust plane high on the south side of Cow Canyon dips 55° west. Approximately 1500 feet below this point, in the bottom of the canyon, the dip decreases to approximately 45°. This decrease in dip, it is believed, continues with depth (Pl. 2, sec. D-D′).

North of the divide at the head of Oak Creek (Pl. 4, fig. 2), Indianola (?) strata have been thrust relatively westward under Precambrian (?) quartzites. Drag folds along this part of the thrust fault are common and relatively large,

PLATE 4.—VIEWS OF CANYON RANGE THRUST

FIGURE 1. VIEW OF EASTERN FRONT OF CANYON RANGE

Showing scarplet center and trace of thrust (dashed line) near crest of range. Precambrian (?) rocks above thrust and Cambrian and younger rocks below.

FIGURE 2. VIEW LOOKING NORTH FROM A POINT ON THE DIVIDE AT HEAD OF OAK CREEK

Precambrian (?) strata dipping west, upper left side of view, are upthrust over Indianola (?) conglomerate, center and right side of view.

FIGURE 1

FIGURE 2

VIEWS OF CANYON RANGE THRUST

VERTICAL AERIAL PHOTOGRAPH SHOWING FAULT PATTERN AT THE SOUTHEAST MARGIN OF THE RANGE

particularly in the footwall block. The general dip of Cretaceous strata in the footwall block east of the thrust fault in this vicinity varies from 10° to 35°. The angle of dip, however, increases rapidly toward the fault zone where locally the Indianola (?) beds are overturned (Pl. 2, sec. D-D′).

South of the divide at the head of Oak Creek, Precambrian (?) strata were thrust into contact with rocks of Cambrian age or younger, and although drag structures have developed, they are apparently not as pronounced as those that occur in the Indianola (?) strata.

South of Dry Fork Canyon, the east and west segments of the thrust fault can be traced without difficulty to the southern end of the range where they disappear beneath Tertiary sediments. North of Dry Fork, however, the exact position of the western segment of the thrust surface is doubtful, and the eastern segment apparently terminates in a eastwest-trending zone of small tear faults and closely spaced joints. North of the head of Bridge Canyon, the thrust fault cuts progressively down through the Cambrian formations into Precambrian (?) rocks. Thus, Precambrian (?) rocks of the thrust sheet were faulted into contact with rocks of approximately the same age, lithology, and structural character, which, in addition to the few and poor exposures in this area, complicated the problem of locating and mapping the thrust fault.

The eastern segment of the thrust fault, between the divide at the head of Oak Creek and the head of Dry Fork and along which Cretaceous rocks were involved in the thrusting, is a local thrust that developed during a second stage of thrusting within the same general zone of an earlier major thrust. This very local ramp-type thrust decreases in displacement from Cow Canyon northward to the vicinity west of the head of Wide Canyon where it terminates in a east-west zone of highly jointed and tear faulted rocks that is mostly covered (Pl. 1).

The most easterly part of the earlier overthrust sheet may be buried beneath Cretaceous and younger sediments that blanket the northeastern and eastern part of the Canyon Range area (Pl. 3). During its development, the thrust sheet may have overridden the earliest Cretaceous sediments, but no supporting evidence has been found.

Tear Faults

Numerous small east-west trending tear faults are present in the northern, intensely folded part of the range between Dry Fork and Leamington Canyon. They are particularly well developed in and north of Wood Canyon. The present sites of most of the canyons draining the western part of the range in this locality have been determined by these tear faults and parallel, closely spaced joint sets. Near the head of Wood Canyon, the Cambrian-Indianola (?) contact has been displaced several hundred feet, and farther down the canyon the Ophir formation has been displaced approximately 800 feet; the block on the south has been displaced relatively eastward.

As these tear faults involved Indianola (?) and older sediments, they most likely developed during the second major diastrophic episode (early Laramide). In the northern part of the range, this orogeny intensified the earlier folds, overturned the west limb of the syncline, and produced the present arcuate pattern of this part of the range. Many of the tear faults were produced by the differential eastward shifting of segments of the arcuate belt of the range and by differential overturning of parts of the west limb of the syncline. There is also another explanation for the development of these tear faults, particularly the small ones and associated parallel joints located immediately north of Dry Fork. Folding was most intense in the northern third of the range. South of Dry Fork, in the central part of the range, folding was less intense and was accompanied by local high-angle thrusting along the eastern front of the range. It is significant that this thrust terminates in the zone of transition between these two parts of the range. Apparently the small tear faults and parallel joints were produced by the differential movement and consequent adjustment that occurred between the central thrust-faulted and northern folded parts of the range. The difference in manner of adjustment to the early Laramide forces in the two parts of the range is discussed under early Laramide orogeny.

Good exposures of vertical to overturned Precambrian (?) and Cambrian strata are present on the south slope of Leamington Canyon. Isolated, small patches are also present on the floor of the canyon and on its northwest slope. On the northwest slope of Leamington Canyon, Precambrian (?) rocks are in fault contact with less intensely deformed Carboniferous strata. The exact nature of this major fault structure is not known, but its parallelism to the smaller tear fault described above indicates that it may be a major transverse or tear fault. Another interpretation is possible. The exact position of the northern extension of the earlier major thrust fault zone is unknown, because the alluvial cover and consequent absence of good exposures has prohibited tracing it northward to Leamington Canyon. The major thrust may emerge from beneath the Precambrian (?) strata and alluvium along the fault zone extending through Leamington Canyon. Such an interpretation would also explain the large stratigraphic throw necessary to bring Carboniferous and Precambrian (?) rocks together.

Normal Faults

Comparatively young border gravity faults occur in a wide zone north and west of Scipio (Sec. F-F', Pl. 2; Pl. 5). The faults nearest the range generally strike north-south. East of the north-south trending faults, and near the central part of Scipio Valley, the faults veer east, varying in strike from N 30° E to approximately N 45° E. The zone that veers to the east may join the southern extension of the normal fault zone along the west front of Long Ridge (not included in area mapped). The north-south fault zone adjacent to the Canyon Range gradually dies out northward and terminates in the vicinity immediately south of the Pass Canyon Mills Road. Facets and apices of very small alluvial fans mark the positions of the faults adjacent to the range. East of the range, on the piedmont slope and in the valley alluvium, their surface expression is in the form of scarplets and sag valleys. Most of the scarplets face the east with the downthrown blocks on the east. East of Scipio, however, they face west and the downthrown blocks are on the west.

Older interior gravity faults occur in the southwest part of the range in a narrow north-south belt between the mouth of Bridge Canyon and the south end of the range. East of Oak City and in the vicinity of the mouth of Whiskey Creek, these faults are well exposed. They are not expressed in the topography, except where there is a difference in the resistance to erosion of the contiguous rocks along the fault. The frontal slope of the range east of Oak City is believed to be a fault-line scarp. Here relatively weak rocks, shale and thin bedded quartzite, were faulted down against massively bedded and relatively resistant quartzite.

Structural History

Early Cretaceous orogeny.—The strata of the Canyon Range area was folded and overriden by a great thrust sheet which apparently moved from west to east. During and after its development, the thrust sheet was deeply eroded and finally buried beneath several thousand feet of Cretaceous sediments (Pl. 3). Several minor pulsations within the area of rising fold and thrust mountains may have caused fluctuations in the processes of weathering and transportation, which probably in part produced the alternating coarse and fine types of sediments that were laid down in the adjacent lowlands. Other features of the sediments, such as the dominance of lenses and extreme particle size variation within short distances, are common.

Spieker (1949, p. 78) described conglomerates possibly of Morrison age which he believes were derived from highlands "not far to the west" and which may represent the debris shed from uplands produced in latest Jurassic time. This either represents a separate and earlier orogeny or an early pulsation of the major early Cretaceous orogeny, as marked by the angular unconformity and coarse Indianola conglomerates.

The Indianola (?) sediments of the Canyon Range, particularly the lower strata, contain coarse conglomerates composed predominantly of Paleozoic limestone and quartzite fragments with subordinate amounts of Mesozoic and Precambrian (?) material. The very coarse character of these sediments indicates that they were shed from mountains of major pro-

portions that developed nearby. The sediments of the Indianola (?) group become coarser westward, indicating that the uplifted area was situated immediately west of the present site of the Canyon Range. The extremely coarse, unsorted, and unrounded boulders and cobbles in the lower part of the sequence could hardly have been derived from sources more than 5 miles away, and probably came from precipitous uplands that varied considerably in composition.

Structural evidence of the early Cretaceous orogeny occurs at many localities in the Canyon Range area. At the end of Yellowstone Canyon, the angular unconformity between the Indianola (?) group and Upper Cambrian limestones and dolomites is well-exposed. Here the conglomerates are upturned and rest on overturned Cambrian strata. In the vicinity of Scipio Peak, the conglomerates rest with angular unconformity on Precambrian (?) rocks, and on the divide between Dry Fork and Wild Horse Canyon, in the axial zone of the synclinal core of the range, the conglomerates rest unconformably on the Ophir formation. The magnitude of the unconformity is clearly shown in these three localities.

The structural relationship of the Indianola (?) conglomerates on the older strata indicates that pre-Indianola (?) folding in the Canyon Range area even including rocks in the thrust sheet, was not intense, since in no locality where the unconformity is exposed is the angular discordance between the Indianola (?) and older strata greater than 20°.

The evidence indicates that, following the stage of major overthrusting, erosion cut deep into the folded and thrusted rocks which, as the basin to the east filled with sediments, overlapped westward onto the thrust sheet and other structures of the Canyon Range area (Pl. 3).

The basal Indianola (?) sediments were deposited over a surface of considerable relief. In the vicinity of Wild Horse Canyon, deposition occurred in a valley with relief of about 800 feet that had developed parallel to the axis of the syncline and in the less resistant Ophir formation. In the vicinity of Scipio Peak, outcrops of Cretaceous conglomerate occur in depressions that represent preserved fragments of subsequent valleys developed on the soft shale and less resistant conglomerate units of the Precambrian (?).

The approximate north-south alignment of the Indianola (?) conglomerate outcrops within the range to the trend of the less resistant rocks on which it was deposited indicates that the relatively gentle folds of the early Cretaceous orogeny were in general parallel to the present folds, subsequent disturbances having only intensified these older folds. The fact that the conglomerate truncates progressively older beds toward the south suggests that the early Cretaceous folds plunged to the north. South of Oak Creek there are no Indianola (?) conglomerate exposures, hence no conclusions can be drawn regarding the altitude of early Cretaceous folds there.

The exact age of this orogeny has not been determined. There is some evidence which indicates that it occurred between the time of deposition of the Navajo sandstone and the beginning of deposition of the Indianola (?) conglomerate, the lower part of which may be Morrison or older. The presence of Navajo sandstone pebbles and cobbles in the Indianola (?) conglomerates, the youngest recognizable rock fragments found in it, and the great amount of erosion that preceded the deposition of the Indianola (?) conglomerates supports the general conclusion that the orogeny began sometime near the close of the Jurassic. The upper time limit is determined by the age of the Indianola (?) conglomerates.

The areal extent of the early Cretaceous orogeny is only partly known. Data now available indicate that it influenced a wide belt from southern Nevada northward through western Utah and eastern Nevada to southern Idaho. Spieker (1946, p. 150–151) describes a Mid-Cretaceous orogeny as having its eastern boundary along the eastern border of the Great Basin province through Utah, eastern Wyoming and veering west in south-central Idaho. Since his writing, additional evidence has been discovered which supports his contention and extends the area of influence of this orogeny westward into Nevada. Thick, highly folded orogenic conglomerates of probable Cretaceous age have been discovered by the writer in Sacramento Pass of the Snake

Mountains, eastern Nevada. Farther west near Eureka, Nevada, MacNeil (1939. p. 335) identified Cretaceous fossils from an unnamed conglomerate that rests unconformably on marine limestone of Carboniferous age. The conglomerate was originally considered Carboniferous. These and other widely scattered outcrops of relatively coarse sediments and orogenic conglomerates of Cretaceous age (Table 3) suggest that the early Cretaceous orogeny influenced a wide belt extending from the Transition zone between the High Plateaus and the Great Basin province to east-central Nevada (Fig. 2). The area involved in the orogeny was evidently a broad, uplifted belt of generally north-south trending mountains and depositional basins.

FIGURE 2. PROBABLE AREAL EXTENT OF THE EARLY CRETACEOUS OROGENY

1. Muddy Mountains, 2. Iron Springs District, 3. Wasatch and Gunnison plateaus, 4. Canyon Range, 5. North central Wasatch, 6. Southeastern Idaho, 7. Snake Mountains, 8. Eureka, Nevada.

The rapidly rising mountains shed large quantities of coarse debris along their margins and into adjacent basins, and a particularly thick blanket of gravels and sands was cast eastward over a piedmont belt in central Utah.

The vast quantity and variety of debris derived from the mountains of this orogenic belt necessitated folding, thrusting, and erosion on a grand scale, which is born out by structural and stratigraphic evidence in and adjacent to

the Canyon Range area. Thrusting during this episode involved the translation of large masses of rock for considerable distances as compared with the later stage of thrusting.

sification of earlier folds, (2) local recurrence of thrusting along old thrust surface, (3) folding of Indianola (?) strata, (4) folding of old thrust surface, (5) tear faulting, particularly in the

TABLE 3.—AGE AND DISTRIBUTION OF COARSE CONGLOMERATES OF MESOZOIC AGE*

	Muddy Mountains [1]	Iron Springs [2]	Wasatch and Gunnison Plateaus [3]	Canyon Range [4]	Central Wasatch [5]	Southeastern Idaho [6]	Snake Mountains, Nev. [7]	Eureka Nevada [8]
Lance								
Montana Colorado	? Overton conglomerate	? Iron Springs formation	Indianola group	? Indianola? group	?	Wayan formation	? Unnamed formation	?
Lower Cretaceous	?	?	May be represented	May be represented	Kelvin Conglomerate	Gannett group	Coarse Conglomerate	Unnamed formation Coarse Conglomerate
Jurassic			Morrison (?)	?	?		?	?

*1. Hewitt (1936, p. 121–122). 2. Mackin (1947, p. 7, 8). 3. Spieker (1949, p. 18–26). 4. This report. 5. Eardley (1944, p. 838). 6. Peck (1941, p. 287). 7. Christiansen, unpublished information. 8. MacNeil (1939, p. 335).

Eardley (1949) has proposed the name Cedar Hills orogeny for this disturbance since the conglomerates and interbedded sandstones and shales in the area of the Cedar Hills reach a maximum known development and represent, according to Spieker (1946, p. 150), a "consecutive response" to a disturbance that probably occurred near the beginning of Upper Cretaceous. Spieker points out, however, that it may have occurred sometime between the "late Jurassic and the early Upper Cretaceous."

Early Laramide orogeny.—The record of this diastrophic episode can be observed in many parts of the Canyon Range and vicinity and consists of several types of structures. These structures developed after the deposition of the Indianola (?) sediments and prior to the deposition of the North Horn (?) formation, the lower part of which may be equivalent to the Price River conglomerates (Spieker, 1946) to the east. This orogeny is correlated with the Early Laramide recognized by Eardley (1944) in the north-central Wasatch Mountains and by Spieker (1949) in the Wasatch Plateau area located east of the Canyon Range.

The kinds of structures produced by this orogeny were largely determined by pre-existing structures. The chief effects were: (1) intensification of earlier folds, (2) local recurrence of thrusting along old thrust surface, (3) folding of Indianola (?) strata, (4) folding of old thrust surface, (5) tear faulting, particularly in the northern part of the range, and (6) local thrusting of Precambrian rocks over the Indianola (?) group.

On the divide between Dry Fork and Wild Horse canyons and within the axial zone of the syncline forming the core of the range, evidence of the intensification of the early Cretaceous folds and folding of the overlying unconformable Indianola (?) beds is clearly shown. Here the Indianola (?) sediments were orignally deposited unconformably on the gently folded and deeply eroded Cambrian strata of the thrust sheet. This deeply eroded open fold and the unconformable overlying Indianola (?) strata were highly deformed. The west limb of the syncline was overturned to the east and the unconformity and overlying Indianola (?) conglomerates were proportionately folded.

North of Wild Horse Canyon, early Laramide folding was intense. In the area of Yellowstone Canyon, the Cambrian-Indianola unconformity was folded to a vertical position and the underlying Cambrian strata were over-turned to the east. Immediately north of Yellowstone Canyon between Wood and the Pass Canyon, folding and overturning was most intense. Here the Tintic quartzite and the Ophir formation are overturned to such an extent that they

approach a recumbent attitude. In this belt of intense overturning, bedding thrusts are common and it is believed that parts of the Middle Cambrian section have been omitted as a result of this thrusting, particularly the incompetent shale and thinly bedded limestone units.

Folding of the old thrust surface and strata in the thrust plate varies considerably in intensity from place to place (Pl. 2,) although in the central and southern parts of the range the intensity of folding was generally less than in the northern part.

North of Bridge Canyon, in the northwest part of the range, the thrust surface is not exposed but is probably overturned to the east. This conclusion is based on the relative degree of overturning of the overlying sedimentary rocks and on the assumption that there was an initial low-angle relationship in this area between the thrust surface and the adjacent strata.

Accompanying the development of these fold structures in the central part of the range were two types of thrust movement: (1) local renewed movement along the old thrust plane and (2) local upthrusting of the Precambrian rocks of the east limb of the syncline, over the Indianola (?) conglomerates. The net displacement of this thrust in the area north of the head of Oak Creek is estimated, from the apparent displacement of the basal beds of the Indianola (?) group and the underlying unconformity, to be approximately 5000 feet. The amount of renewed movement on the western segment of the thrust could not be determined, but was probably equal to or perhaps locally in excess of that which occurred on the thrust involving the Indianola (?) conglomerate. Renewed movement along the west segment of the fault in the vicinity of Oak Creek was relatively up and westward and in a reverse direction from that of the earlier overthrusting which was from west to east. The reversal in direction of movement is suggested by the chaotic nature of the drag structures associated with the western segment of the thrust. The drag structures of the eastern segment, however, where renewed movement was in the same relative direction as the earlier overthrust movement, are relatively simple and normal. It appears that the present synclinal core of the range, composed of comparatively competent Precambrian strata of the thrust sheet, was folded, and during folding was thrust relatively up and out along the older thrust plane over adjacent younger rocks to the east and the west, in the same general manner that higher beds in flexure folding move relatively up and out over the lower beds.

The early Laramide high-angle ramp-type thrust north of the head of Oak Creek, which involved the Indianola (?) conglomerate, diminishes in amount of displacement northward to a point near the head of Dry Fork where it apparently terminates in a series of very small east-west, nearly vertical tear faults and closely spaced joints that parallel the tear faults. It is significant that the fault terminates in the area where the nearly vertical unconformity beneath the Indianola (?) sediments veers westward toward the axial zone of the major syncline. Pre-Indianola (?) erosion north of the point of termination of the thrust completely removed the east limb of the syncline, consequently the pre-early Laramide structural composition of the north part of the range was considerably different from the central and southern parts. This difference in structural composition between the northern and central parts of the range is believed responsible for the manner in which they failed under the influence of the early Laramide forces and further explains the termination of the thrust fault. The part of the range north of Dry Fork and north of the termination of the thrust fault yielded to the early Laramide forces mainly by intensification of the earlier folds. Precambrian rocks were thrust over the Indianola (?) sediments in the east-central part of the range, where the relatively competent strata of the east limb of the syncline were preserved and buried beneath the younger sediments. Consequently this competent limb acted as a strut athwart the compressional stresses of the early Laramide forces; it was not folded as was the northern part of the range, where the competent limb had been removed by Pre-Indianola (?) erosion, but was thrust up through the overlying Cretaceous sediments. The east-west zone of small tear faults and joints at the northern terminus of the thrust fault developed as a result of the differential movement between the folded segment and locally thrusted segment of the range.

Post-North Horn mild folding.—In the northeast piedmont area of the Canyon Range, the North Horn formation rests unconformably on the upper strata of the Indianola (?) group. The angular relationship between these formations is low, from 4° to about 8°. Unconformably above the North Horn formation is the Fool Creek conglomerate of Oligocene (?) age. The angular discordance between the Fool Creek conglomerate and the North Horn formation ranges up to a maximum of about 20°.

This structural relationship indicates that, after deposition of the North Horn formation, a mild disturbance tilted and mildly folded the North Horn and older formations and produced essentially eastward-dipping homoclinal structures in the northeastern piedmont area. The intensity of deformation is suggested by the angles of dip of the North Horn formation which range from 15° to about 25°.

The amount of deformation produced within the range by this disturbance is not known since the relatively young structures and formations are confined to the piedmont areas. Probably, however, the main block of the range was broadly upwarped in general conformity with the disturbance of the North Horn formation.

This disturbance could not be dated precisely because of the relatively long time interval between the limiting structures and formations. However, it may correlate with one of the major events of the Wasatch Plateau which flexed and faulted the Wasatch monocline sometime between late Eocene and the Miocene, according to Spieker (1949, p. 80). Further tentative correlation is made with a stage of gentle folding of Watsatch strata of the north-central Wasatch Mountains described by Eardley (1944, p. 862).

Post-Fool Creek disturbance.—Evidence of a post-Fool Creek disturbance is meager and inconclusive. In the foothill belt of the southwest part of the range south of Oak City, there are several isolated bed-rock exposures of the Fool Creek conglomerate whose beds are tilted and disturbed. The strata of the conglomerate in this area dip westward as much as 40°. The outcrops displaying relatively high angles of dip are local and therefore difficult to evaluate in terms of a general widespread disturbance. The strata of the Fool Creek conglomerate within the range are essentially undisturbed. The fact that the disturbed Fool Creek strata occur in outcrops along the western margin of the range, where older normal faults are present, suggests that the tilting may be associated with movement along the older normal faults.

Early normal faulting.—Relatively old normal faults occur within the mountain block. They do not express themselves topographically except as the result of subsequent erosion. Two relatively prominent ones occur in the southwest part of the range. (Pl. 2, Sec. E-E′) The maximum displacement on the eastern fault is about 2500 feet and on the western fault about 450 feet. They trend approximately north-south and north of Oak Creek; the larger eastern fault parallels the western front of the range for several miles near its western base. It is believed that the fault in this vicinity is indirectly responsible for the development of the western frontal slope of the range and that this frontal slope is a fault-line scarp. The Ophir formation and younger sedimentary rocks were here faulted down against the more resistant Precambrian (?) quartzites, producing ideal conditions for differential erosion and the development of the fault line scarp.

The faults are definitely older than the border normal faults which produced part of the present relief of the range. There is a possibility, but no definite evidence, that they pre-date the Fool Creek conglomerate. Similar normal faults have been described by Nolan (1935), Calkins and Butler (1943), Longwell (1926), and Spieker (1949) in areas adjacent to the Canyon Range, but available evidence is insufficient to make definite correlation.

Basin and range faulting.—Structural and physiographic evidence indicates that part of the present relief of the range is the direct result of basin and range normal faulting and that the mountain block has been slightly tilted to the west and probably also tilted to the north. Evidence of basin and range normal faulting is confined to the foothills belt and valley at the southeastern margin of the range. Within this belt, the following kinds of evidence of normal faulting are present: (1) scarplets, (2) abnormally small alluvial fans, (3) sag structures, (4) abrupt and imposing east facing

Good exposures of vertical to overturned Precambrian (?) and Cambrian strata are present on the south slope of Leamington Canyon. Isolated, small patches are also present on the floor of the canyon and on its northwest slope. On the northwest slope of Leamington Canyon, Precambrian (?) rocks are in fault contact with less intensely deformed Carboniferous strata. The exact nature of this major fault structure is not known, but its parallelism to the smaller tear fault described above indicates that it may be a major transverse or tear fault. Another interpretation is possible. The exact position of the northern extension of the earlier major thrust fault zone is unknown, because the alluvial cover and consequent absence of good exposures has prohibited tracing it northward to Leamington Canyon. The major thrust may emerge from beneath the Precambrian (?) strata and alluvium along the fault zone extending through Leamington Canyon. Such an interpretation would also explain the large stratigraphic throw necessary to bring Carboniferous and Precambrian (?) rocks together.

Normal Faults

Comparatively young border gravity faults occur in a wide zone north and west of Scipio (Sec. F-F', Pl. 2; Pl. 5). The faults nearest the range generally strike north-south. East of the north-south trending faults, and near the central part of Scipio Valley, the faults veer east, varying in strike from N 30° E to approximately N 45° E. The zone that veers to the east may join the southern extension of the normal fault zone along the west front of Long Ridge (not included in area mapped). The north-south fault zone adjacent to the Canyon Range gradually dies out northward and terminates in the vicinity immediately south of the Pass Canyon Mills Road. Facets and apices of very small alluvial fans mark the positions of the faults adjacent to the range. East of the range, on the piedmont slope and in the valley alluvium, their surface expression is in the form of scarplets and sag valleys. Most of the scarplets face the east with the downthrown blocks on the east. East of Scipio, however, they face west and the downthrown blocks are on the west.

Older interior gravity faults occur in the southwest part of the range in a narrow north-south belt between the mouth of Bridge Canyon and the south end of the range. East of Oak City and in the vicinity of the mouth of Whiskey Creek, these faults are well exposed. They are not expressed in the topography, except where there is a difference in the resistance to erosion of the contiguous rocks along the fault. The frontal slope of the range east of Oak City is believed to be a fault-line scarp. Here relatively weak rocks, shale and thin bedded quartzite, were faulted down against massively bedded and relatively resistant quartzite.

Structural History

Early Cretaceous orogeny.—The strata of the Canyon Range area was folded and overriden by a great thrust sheet which apparently moved from west to east. During and after its development, the thrust sheet was deeply eroded and finally buried beneath several thousand feet of Cretaceous sediments (Pl. 3). Several minor pulsations within the area of rising fold and thrust mountains may have caused fluctuations in the processes of weathering and transportation, which probably in part produced the alternating coarse and fine types of sediments that were laid down in the adjacent lowlands. Other features of the sediments, such as the dominance of lenses and extreme particle size variation within short distances, are common.

Spieker (1949, p. 78) described conglomerates possibly of Morrison age which he believes were derived from highlands "not far to the west" and which may represent the debris shed from uplands produced in latest Jurassic time. This either represents a separate and earlier orogeny or an early pulsation of the major early Cretaceous orogeny, as marked by the angular unconformity and coarse Indianola conglomerates.

The Indianola (?) sediments of the Canyon Range, particularly the lower strata, contain coarse conglomerates composed predominantly of Paleozoic limestone and quartzite fragments with subordinate amounts of Mesozoic and Precambrian (?) material. The very coarse character of these sediments indicates that they were shed from mountains of major pro-

portions that developed nearby. The sediments of the Indianola (?) group become coarser westward, indicating that the uplifted area was situated immediately west of the present site of the Canyon Range. The extremely coarse, unsorted, and unrounded boulders and cobbles in the lower part of the sequence could hardly have been derived from sources more than 5 miles away, and probably came from precipitous uplands that varied considerably in composition.

Structural evidence of the early Cretaceous orogeny occurs at many localities in the Canyon Range area. At the end of Yellowstone Canyon, the angular unconformity between the Indianola (?) group and Upper Cambrian limestones and dolomites is well-exposed. Here the conglomerates are upturned and rest on overturned Cambrian strata. In the vicinity of Scipio Peak, the conglomerates rest with angular unconformity on Precambrian (?) rocks, and on the divide between Dry Fork and Wild Horse Canyon, in the axial zone of the synclinal core of the range, the conglomerates rest unconformably on the Ophir formation. The magnitude of the unconformity is clearly shown in these three localities.

The structural relationship of the Indianola (?) conglomerates on the older strata indicates that pre-Indianola (?) folding in the Canyon Range area even including rocks in the thrust sheet, was not intense, since in no locality where the unconformity is exposed is the angular discordance between the Indianola (?) and older strata greater than 20°.

The evidence indicates that, following the stage of major overthrusting, erosion cut deep into the folded and thrusted rocks which, as the basin to the east filled with sediments, overlapped westward onto the thrust sheet and other structures of the Canyon Range area (Pl. 3).

The basal Indianola (?) sediments were deposited over a surface of considerable relief. In the vicinity of Wild Horse Canyon, deposition occurred in a valley with relief of about 800 feet that had developed parallel to the axis of the syncline and in the less resistant Ophir formation. In the vicinity of Scipio Peak, outcrops of Cretaceous conglomerate occur in depressions that represent preserved fragments of subsequent valleys developed on the soft shale and less resistant conglomerate units of the Precambrian (?).

The approximate north-south alignment of the Indianola (?) conglomerate outcrops within the range to the trend of the less resistant rocks on which it was deposited indicates that the relatively gentle folds of the early Cretaceous orogeny were in general parallel to the present folds, subsequent disturbances having only intensified these older folds. The fact that the conglomerate truncates progressively older beds toward the south suggests that the early Cretaceous folds plunged to the north. South of Oak Creek there are no Indianola (?) conglomerate exposures, hence no conclusions can be drawn regarding the altitude of early Cretaceous folds there.

The exact age of this orogeny has not been determined. There is some evidence which indicates that it occurred between the time of deposition of the Navajo sandstone and the beginning of deposition of the Indianola (?) conglomerate, the lower part of which may be Morrison or older. The presence of Navajo sandstone pebbles and cobbles in the Indianola (?) conglomerates, the youngest recognizable rock fragments found in it, and the great amount of erosion that preceded the deposition of the Indianola (?) conglomerates supports the general conclusion that the orogeny began sometime near the close of the Jurassic. The upper time limit is determined by the age of the Indianola (?) conglomerates.

The areal extent of the early Cretaceous orogeny is only partly known. Data now available indicate that it influenced a wide belt from southern Nevada northward through western Utah and eastern Nevada to southern Idaho. Spieker (1946, p. 150–151) describes a Mid-Cretaceous orogeny as having its eastern boundary along the eastern border of the Great Basin province through Utah, eastern Wyoming and veering west in south-central Idaho. Since his writing, additional evidence has been discovered which supports his contention and extends the area of influence of this orogeny westward into Nevada. Thick, highly folded orogenic conglomerates of probable Cretaceous age have been discovered by the writer in Sacramento Pass of the Snake

Mountains, eastern Nevada. Farther west near Eureka, Nevada, MacNeil (1939. p. 335) identified Cretaceous fossils from an unnamed conglomerate that rests unconformably on marine limestone of Carboniferous age. The conglomerate was originally considered Carboniferous. These and other widely scattered outcrops of relatively coarse sediments and orogenic conglomerates of Cretaceous age (Table 3) suggest that the early Cretaceous orogeny influenced a wide belt extending from the Transition zone between the High Plateaus and the Great Basin province to east-central Nevada (Fig. 2). The area involved in the orogeny was evidently a broad, uplifted belt of generally north-south trending mountains and depositional basins.

FIGURE 2. PROBABLE AREAL EXTENT OF THE EARLY CRETACEOUS OROGENY

1. Muddy Mountains, 2. Iron Springs District, 3. Wasatch and Gunnison plateaus, 4. Canyon Range, 5. North central Wasatch, 6. Southeastern Idaho, 7. Snake Mountains, 8. Eureka, Nevada.

The rapidly rising mountains shed large quantities of coarse debris along their margins and into adjacent basins, and a particularly thick blanket of gravels and sands was cast eastward over a piedmont belt in central Utah.

The vast quantity and variety of debris derived from the mountains of this orogenic belt necessitated folding, thrusting, and erosion on a grand scale, which is born out by structural and stratigraphic evidence in and adjacent to

the Canyon Range area. Thrusting during this episode involved the translation of large masses of rock for considerable distances as compared with the later stage of thrusting.

sification of earlier folds, (2) local recurrence of thrusting along old thrust surface, (3) folding of Indianola (?) strata, (4) folding of old thrust surface, (5) tear faulting, particularly in the

TABLE 3.—AGE AND DISTRIBUTION OF COARSE CONGLOMERATES OF MESOZOIC AGE*

	Muddy Mountains[1]	Iron Springs[2]	Wasatch and Gunnison Plateaus[3]	Canyon Range[4]	Central Wasatch[5]	Southeastern Idaho[6]	Snake Mountains, Nev.[7]	Eureka Nevada[8]
Lance								
Montana Colorado	? ↑ Overton conglomerate	? ↑ Iron Springs formation	Indianola group	? ↑ Indianola ? group	? ↑	Wayan formation	? ↑ Unnamed formation	? ↑
Lower Cretaceous	↓ ?	↓ ?	May be represented	May be represented	Kelvin Conglomerate	Gannett group	Coarse Conglomerate	Unnamed formation Coarse Conglomerate
Jurassic			Morrison (?)	↓ ?	↓ ?		↓ ?	↓ ?

*1. Hewitt (1936, p. 121–122). 2. Mackin (1947, p. 7, 8). 3. Spieker (1949, p. 18–26). 4. This report. 5. Eardley (1944, p. 838). 6. Peck (1941, p. 287). 7. Christiansen, unpublished information. 8. MacNeil (1939, p. 335).

Eardley (1949) has proposed the name Cedar Hills orogeny for this disturbance since the conglomerates and interbedded sandstones and shales in the area of the Cedar Hills reach a maximum known development and represent, according to Spieker (1946, p. 150), a "consecutive response" to a disturbance that probably occurred near the beginning of Upper Cretaceous. Spieker points out, however, that it may have occurred sometime between the "late Jurassic and the early Upper Cretaceous."

Early Laramide orogeny.—The record of this diastrophic episode can be observed in many parts of the Canyon Range and vicinity and consists of several types of structures. These structures developed after the deposition of the Indianola (?) sediments and prior to the deposition of the North Horn (?) formation, the lower part of which may be equivalent to the Price River conglomerates (Spieker, 1946) to the east. This orogeny is correlated with the Early Laramide recognized by Eardley (1944) in the north-central Wasatch Mountains and by Spieker (1949) in the Wasatch Plateau area located east of the Canyon Range.

The kinds of structures produced by this orogeny were largely determined by pre-existing structures. The chief effects were: (1) intensification of earlier folds, (2) local recurrence of thrusting along old thrust surface, (3) folding of Indianola (?) strata, (4) folding of old thrust surface, (5) tear faulting, particularly in the northern part of the range, and (6) local thrusting of Precambrian rocks over the Indianola (?) group.

On the divide between Dry Fork and Wild Horse canyons and within the axial zone of the syncline forming the core of the range, evidence of the intensification of the early Cretaceous folds and folding of the overlying unconformable Indianola (?) beds is clearly shown. Here the Indianola (?) sediments were orignally deposited unconformably on the gently folded and deeply eroded Cambrian strata of the thrust sheet. This deeply eroded open fold and the unconformable overlying Indianola (?) strata were highly deformed. The west limb of the syncline was overturned to the east and the unconformity and overlying Indianola (?) conglomerates were proportionately folded.

North of Wild Horse Canyon, early Laramide folding was intense. In the area of Yellowstone Canyon, the Cambrian-Indianola unconformity was folded to a vertical position and the underlying Cambrian strata were over-turned to the east. Immediately north of Yellowstone Canyon between Wood and the Pass Canyon, folding and overturning was most intense. Here the Tintic quartzite and the Ophir formation are overturned to such an extent that they

approach a recumbent attitude. In this belt of intense overturning, bedding thrusts are common and it is believed that parts of the Middle Cambrian section have been omitted as a result of this thrusting, particularly the incompetent shale and thinly bedded limestone units.

Folding of the old thrust surface and strata in the thrust plate varies considerably in intensity from place to place (Pl. 2,) although in the central and southern parts of the range the intensity of folding was generally less than in the northern part.

North of Bridge Canyon, in the northwest part of the range, the thrust surface is not exposed but is probably overturned to the east. This conclusion is based on the relative degree of overturning of the overlying sedimentary rocks and on the assumption that there was an initial low-angle relationship in this area between the thrust surface and the adjacent strata.

Accompanying the development of these fold structures in the central part of the range were two types of thrust movement: (1) local renewed movement along the old thrust plane and (2) local upthrusting of the Precambrian rocks of the east limb of the syncline, over the Indianola (?) conglomerates. The net displacement of this thrust in the area north of the head of Oak Creek is estimated, from the apparent displacement of the basal beds of the Indianola (?) group and the underlying unconformity, to be approximately 5000 feet. The amount of renewed movement on the western segment of the thrust could not be determined, but was probably equal to or perhaps locally in excess of that which occurred on the thrust involving the Indianola (?) conglomerate. Renewed movement along the west segment of the fault in the vicinity of Oak Creek was relatively up and westward and in a reverse direction from that of the earlier overthrusting which was from west to east. The reversal in direction of movement is suggested by the chaotic nature of the drag structures associated with the western segment of the thrust. The drag structures of the eastern segment, however, where renewed movement was in the same relative direction as the earlier overthrust movement, are relatively simple and normal. It appears that the present synclinal core of the range, composed of comparatively competent Precambrian strata of the thrust sheet, was folded, and during folding was thrust relatively up and out along the older thrust plane over adjacent younger rocks to the east and the west, in the same general manner that higher beds in flexure folding move relatively up and out over the lower beds.

The early Laramide high-angle ramp-type thrust north of the head of Oak Creek, which involved the Indianola (?) conglomerate, diminishes in amount of displacement northward to a point near the head of Dry Fork where it apparently terminates in a series of very small east-west, nearly vertical tear faults and closely spaced joints that parallel the tear faults. It is significant that the fault terminates in the area where the nearly vertical unconformity beneath the Indianola (?) sediments veers westward toward the axial zone of the major syncline. Pre-Indianola (?) erosion north of the point of termination of the thrust completely removed the east limb of the syncline, consequently the pre-early Laramide structural composition of the north part of the range was considerably different from the central and southern parts. This difference in structural composition between the northern and central parts of the range is believed responsible for the manner in which they failed under the influence of the early Laramide forces and further explains the termination of the thrust fault. The part of the range north of Dry Fork and north of the termination of the thrust fault yielded to the early Laramide forces mainly by intensification of the earlier folds. Precambrian rocks were thrust over the Indianola (?) sediments in the east-central part of the range, where the relatively competent strata of the east limb of the syncline were preserved and buried beneath the younger sediments. Consequently this competent limb acted as a strut athwart the compressional stresses of the early Laramide forces; it was not folded as was the northern part of the range, where the competent limb had been removed by Pre-Indianola (?) erosion, but was thrust up through the overlying Cretaceous sediments. The east-west zone of small tear faults and joints at the northern terminus of the thrust fault developed as a result of the differential movement between the folded segment and locally thrusted segment of the range.

Post-North Horn mild folding.—In the northeast piedmont area of the Canyon Range, the North Horn formation rests unconformably on the upper strata of the Indianola (?) group. The angular relationship between these formations is low, from 4° to about 8°. Unconformably above the North Horn formation is the Fool Creek conglomerate of Oligocene (?) age. The angular discordance between the Fool Creek conglomerate and the North Horn formation ranges up to a maximum of about 20°.

This structural relationship indicates that, after deposition of the North Horn formation, a mild disturbance tilted and mildly folded the North Horn and older formations and produced essentially eastward-dipping homoclinal structures in the northeastern piedmont area. The intensity of deformation is suggested by the angles of dip of the North Horn formation which range from 15° to about 25°.

The amount of deformation produced within the range by this disturbance is not known since the relatively young structures and formations are confined to the piedmont areas. Probably, however, the main block of the range was broadly upwarped in general conformity with the disturbance of the North Horn formation.

This disturbance could not be dated precisely because of the relatively long time interval between the limiting structures and formations. However, it may correlate with one of the major events of the Wasatch Plateau which flexed and faulted the Wasatch monocline sometime between late Eocene and the Miocene, according to Spieker (1949, p. 80). Further tentative correlation is made with a stage of gentle folding of Watsatch strata of the north-central Wasatch Mountains described by Eardley (1944, p. 862).

Post-Fool Creek disturbance.—Evidence of a post-Fool Creek disturbance is meager and inconclusive. In the foothill belt of the southwest part of the range south of Oak City, there are several isolated bed-rock exposures of the Fool Creek conglomerate whose beds are tilted and disturbed. The strata of the conglomerate in this area dip westward as much as 40°. The outcrops displaying relatively high angles of dip are local and therefore difficult to evaluate in terms of a general widespread disturbance. The strata of the Fool Creek conglomerate within the range are essentially undisturbed. The fact that the disturbed Fool Creek strata occur in outcrops along the western margin of the range, where older normal faults are present, suggests that the tilting may be associated with movement along the older normal faults.

Early normal faulting.—Relatively old normal faults occur within the mountain block. They do not express themselves topographically except as the result of subsequent erosion. Two relatively prominent ones occur in the southwest part of the range. (Pl. 2, Sec. E-E′) The maximum displacement on the eastern fault is about 2500 feet and on the western fault about 450 feet. They trend approximately north-south and north of Oak Creek; the larger eastern fault parallels the western front of the range for several miles near its western base. It is believed that the fault in this vicinity is indirectly responsible for the development of the western frontal slope of the range and that this frontal slope is a fault-line scarp. The Ophir formation and younger sedimentary rocks were here faulted down against the more resistant Precambrian (?) quartzites, producing ideal conditions for differential erosion and the development of the fault line scarp.

The faults are definitely older than the border normal faults which produced part of the present relief of the range. There is a possibility, but no definite evidence, that they pre-date the Fool Creek conglomerate. Similar normal faults have been described by Nolan (1935), Calkins and Butler (1943), Longwell (1926), and Spieker (1949) in areas adjacent to the Canyon Range, but available evidence is insufficient to make definite correlation.

Basin and range faulting.—Structural and physiographic evidence indicates that part of the present relief of the range is the direct result of basin and range normal faulting and that the mountain block has been slightly tilted to the west and probably also tilted to the north. Evidence of basin and range normal faulting is confined to the foothills belt and valley at the southeastern margin of the range. Within this belt, the following kinds of evidence of normal faulting are present: (1) scarplets, (2) abnormally small alluvial fans, (3) sag structures, (4) abrupt and imposing east facing

mountain front, (5) triangular facets, (6) alignment of facets, and (7) asymmetrical range with crest line near the eastern front and near the fault zone.

Numerous scarplets are present within the alluviated sag valley in which Scipio is located (Pl. 5). Most of these face east, but east of Scipio, the scarplets face west, indicating that the lowest part of the Scipio Valley is essentially a graben structure. The general trend of the zone of scarplets is north-south, but, several scarplets north and northwest of Scipio veer to the east. The most prominent scarplet in the valley, about 2 miles north-west of Scipio, changes its trend abruptly from approximately north-south to north 45° east. The approximate vertical displacement is greatest at the point of change in strike, and aggregates about 125 feet. The total displacement resulted from two distinct movements, the first of about 115 feet and the second of about 10 feet. After the first movement, the downthrown block was partly alluviated and the streams established new profiles of equilibrium after eroding small but comparatively wide flat-bottomed valleys on the upthrown side. The second movement displaced these local surfaces of steam equilibrium approximately 10 feet and produced fresh surfaces on the older escarpment. Following the second movement, the intermittent streams gullied the small valley bottoms on the upthrown block and superimposed very small alluvial fans on the older ones of the downthrown side. This small-scale fan-on-fan structure also characterizes the larger alluvial deposits near the range. There are several generations of superimposed fans at the mouths of essentially all the canyons that drain the southeast part of the range. At the mouths of Little Long Canyon and John Williams Canyon, this fan-on-fan structure is particularly well-developed. The successively lower gradients of the exposed portions of successively older and larger alluvial fans suggest that the valley block has rotated to the west with each movement.

Small sag or rift structures without surface outlet drainage are adjacent to many of the scarplets. These minor structures occur within a much larger sag or structural basin, the Scipio Valley, which, like the smaller ones, has no surface drainage outlet, and forms a prominent structural depression within the zone of normal faulting. The larger basin is about 7 miles wide and 10 miles long and probably represents a widened and relatively downwarped portion of the floor of a graben that extends north and south of Scipio Valley.

The abrupt and imposing southeastern front of the range and the presence of aligned triangular facets, though not diagnostic, indicate normal faulting. The asymmetrical nature of the range is further evidence of normal faulting and tilting of the mountain bock. In the southern part of the range, the westward-draining streams have comparatively low gradients and range from 8 to 10 miles long. In this same area, the gradients of the eastward draining streams are extremely steep, usually heading in vertical ledges, and are 1–2 miles long. This asymmetry, it is believed, was produced by westward tilting of the range as normal faulting progressed.

The highest points of elevation are located in the southeast part of the range. Elevations in general decrease northward and westward. The northward decrease in elevation is also true of the broad east-west gravel and conglomerate-filled divides along the crest of the range which prior to Basin and Range faulting were probably near the same elevation. There is also a close correlation between the number of normal faults in the eastern piedmont normal fault belt and the height of the range. The maximum number faults occur opposite the highest elevations. This suggests that the range was tilted to the north as well as to the west.

In the east-central part of the range, the strata of the Fool Creek conglomerate were broken by Basin and Range faulting, but individual beds or units could not be correlated across the fault. Therefore, in this locality, the throw of the fault nearest the range can only be estimated. The difference in elevation of the upper surfaces of the Fool Creek conglomerates in the hanging wall and footwall blocks is approximately 1800 feet. If it is assumed that the upper surfaces of the Fool Creek conglomerate prior to being displaced were at the same elevation and that they have been eroded equally since faulting, the throw of the fault would be approximately 1800 feet. South of this area, the aggregate displacement of the several

faults may reach 3500 feet, although evidence supporting this figure was not discovered.

Since its inception, Basin and Range faulting has continued intermittently to the present, as indicated by the structural and physiographic features of the area. The precise time of initiation of faulting could not be determined from the known facts, but it probably began in late Miocene or early Pliocene time.

Summary

The major exposed structural features recorded in the rocks of the Canyon Range developed during two major stages of deformation. Several other less intense disturbances are indicated. The following is a summary of the stages that have been recognized and their general nature, beginning with the oldest:

1. Early Cretaceous orogeny—The first major episode of diastrophism recorded in the rocks of the Canyon Range probably began in late Jurassic time and culminated in the Lower Cretaceous. It involved uplift, folding, and overthrusting from the west on a large scale. The thrust plate or sheet has a minimum thickness of 12,000 feet and has been displaced, in all probability, not less than 10 miles.

2. Early Laramide orogeny—In late Colorado time the second major diastrophic event was recorded in the rocks of the Canyon Range. The following is a summary of the kinds of adjustment that occurred: (a) older folds were intensified; (b) Indianola (?) conglomerates were folded; (c) the older thrust plane was folded and locally renewed movement occurred along it; and (d) the eastern limb of the synclinal core of the range was locally thrust up and out over the Indianola (?) conglomerate. The compressional forces of this diastrophic event produced high-angle, ramp-type thrusts in the area north of the head of Oak Creek where Cretaceous rocks were involved in the thrusting.

3. Post-North Horn mild folding—A third and less intense disturbance followed the deposition of the North Horn formation. It produced mild folds and in general tilted the North Horn strata to the east, producing essentially homoclinal structures in the northeast piedmont of the range.

4. Post-Fool Creek disturbance—In the southwest foothills of the Canyon Range, the strata of the Fool Creek conglomerate are locally tilted westward 30° to 40°. The extent and consequently the significance of this disturbance is not known since outcrops are very limited. The tilted Fool Creek strata may be related to the early stage of normal faulting.

5. Early normal faulting—Comparatively old normal faults occur in the southwest interior part of the range. They are older than the border normal faults and may predate the Fool Creek conglomerate, although definite evidence for this relationship was not discovered.

6. Basin and Range faulting—The present physiographic features of the Canyon Range developed largely as a result of normal faulting that probably began in late Miocene time and has continued to the Recent. Fault scarps, facets, sag valleys, abnormally small alluvial fans, and the asymmetry of the range indicate that it is a fault block.

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University of Utah, Salt Lake City, Utah.
Manuscript Received by the Secretary of the Society, September 24, 1951.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 63, PP. 741-744 JULY 1952

Short Notes

REEF FORMATION IN THE GULF OF MEXICO OFF APALACHICOLA BAY, FLORIDA

BY G. F. JORDAN

Until recent years the continental slope extending far out into the Gulf of Mexico was generally considered to be a great submerged plain, tilted, but unmarked by any significant topographic relief. This misconception was based on relatively few soundings spaced far apart on the early reconnaissance surveys of 75 to 100 years ago. Modern detailed surveys have gradually pulled the veil away from this submerged land, uncovering numerous interesting topographic features.

One of the first significant submarine features uncovered was the ancient valley of the Mississippi River. This valley, indenting the continental slope for a distance of about 80 miles, lies about 35 miles southwest of the present delta. It is approximately 20 miles wide and 1000 feet deep where its floor blends with the continental slope. The head of another valley, an expansive submarine embayment, lies only 15 miles off the Florida coast line at Pensacola. Farther offshore in this area in 300-fathom depths is De Soto Canyon, a distinctive sediment-filled trough 25 miles long, 2 miles wide, and more than 600 feet deep.

Smaller features in the form of hills, domes, and coral heads dot the offshore slope in a direction which parallels the coast line of Louisiana and Texas. A large number of these features lie in depths greater than 50 fathoms. Paralleling the southern Texas coast in 40 to 50 fathoms, numerous groups of coral heads rise to 25- to 45-fathom depths. In one of these groups all the heads are at the same depth. Also, groups of coral heads lie off the Rio Grande and Laguna Madre at about the same depth—some at 18 fathoms, others nearer the coast at 14 fathoms.

The writer (Jordan, 1951) has recently described several interesting features on Florida's continental slope in an area covering 25,000 square miles where depths range from 30 to 1750 fathoms. The most outstanding feature in this area is the continental escarpment beginning in 700- to 900-fathom depths and dropping 5000 feet in a distance of 2 miles. In this same article the Continental Shelf Margin is clearly illustrated by fathometer profiles which show a definite break in 40-fathom depths.

Hydrographic surveys have now been extended eastward, nearer the Florida Peninsula. The continental shelf covered by these surveys has little relief and is in general undulating. There is one significant exception, however, in an area 70 miles south-southwest of Apalachicola Bay, 100 miles northwest of Tampa Bay, Florida (Fig. 1). A distinctive reef formation was found here, extending over 300 square miles in depths of 18 to 30 fathoms.

A reconnaissance survey of the area was first made by the Coast and Geodetic Survey in 1914 when shoal depths were reported by fishermen. That survey revealed reefs at depths of 14 fathoms in general depths of 20 to 25 fathoms. The submerged reefs have now been surveyed in detail with modern electronic equipment. Numerous continuous profiles of the bottom were obtained by recording fathometers while the survey vessel's position was accurately fixed at all times by positioning equipment electronically tied to control points on land.

This modern complete survey provided an abundance of hydrographic data from which an accurate bathymetric chart could be constructed. By contouring the area at 1-fathom intervals a distinctive formation is brought into relief (Fig. 2). It has the appearance of an old river delta. The 3-mile wide "river bed" has a north and south axis, and it is flanked by embankments 50 feet high. The seaward end, or "mouth," is divided by an extensive shoal or reef area, and small reefs dot the "passes."

Paralleling the existence of this feature dis-

tinctive on Florida's Continental Shelf in the Gulf of Mexico are other physiographic details which should prove of interest to geologists and oceanographers. The bottom samples obtained

time is well known. Old shore lines cut by waves during the glacial period were subsequently drowned by the rising sea. Perhaps Florida Middle Ground also gives evidence which will

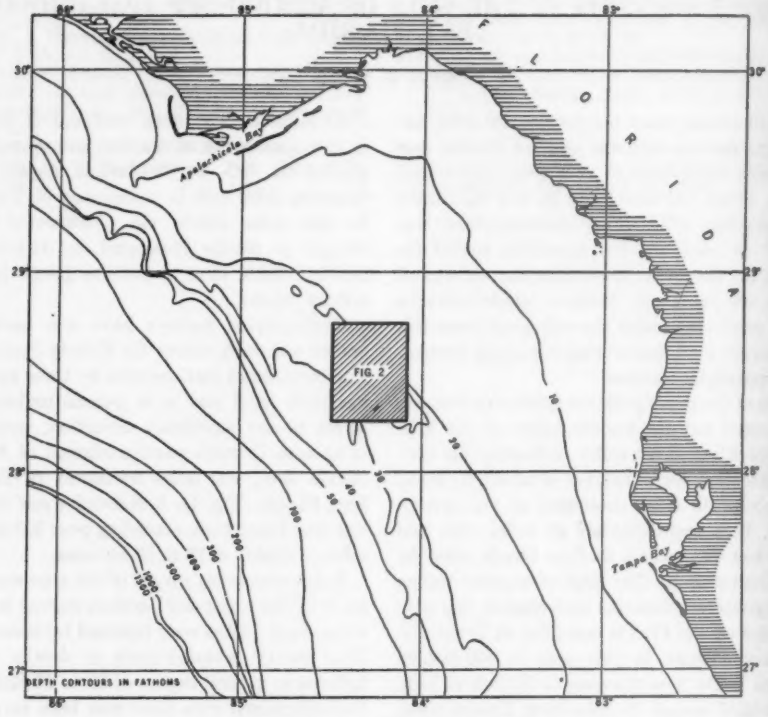


FIGURE 1. INDEX MAP

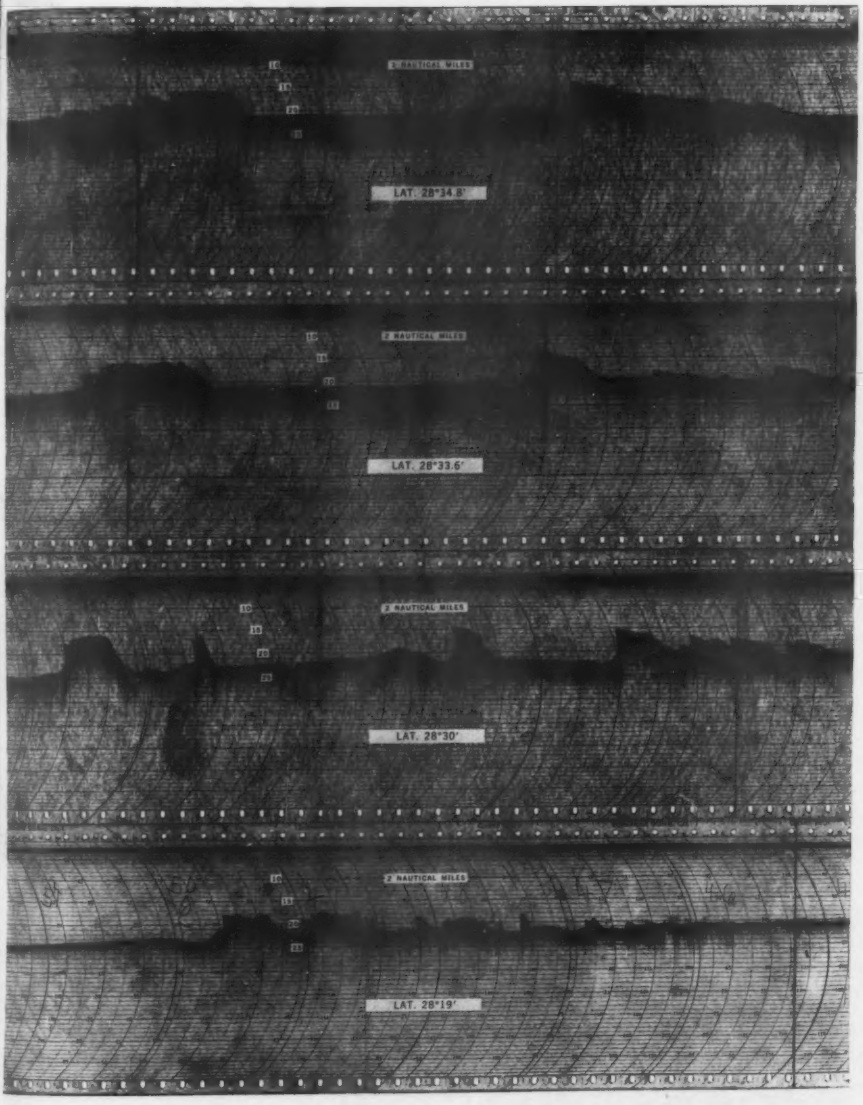
on top of the embankments and shoals, and the character of the bottom revealed by the recorded profiles, indicate that all or most of the reefs are topped with coral. That the coral heads lie in the same depth plane is a circumstance which occurs in other parts of the Gulf. Also, the numerous shoals and depressions at approximately 22-fathom depth and the coincidence of a bottom gradient change at 22 fathoms must have some significance. The terrace inshore from this area is on a 1 foot per mile gradient, contrasting with a slope of 6 feet per mile in adjacent seaward depths.

That sea level has fluctuated in past geologic

help to reconstruct the geological history of Florida and the Gulf of Mexico.

Four sections of graphic profiles recorded by fathometer are illustrated in Plate 1. The bottom gradients are greatly exaggerated on these records because the usefulness and efficiency of the instrumentation require a contracted horizontal scale, and this results in an exaggeration of the vertical scale. Although the slopes on some of the reefs appear to be nearly vertical, the maximum slope is less than 11°. This slope, however, is unusual in these continental-shelf waters.

This area has a special significance in the



FATHOGRAM PROFILES

Vertical scales exaggerated 55 to 60 times, depending on the speed of the survey vessel. Depths are in fathoms.

FIGURE 2. FLORIDA MIDDLE GROUND

fishing industry. The irregular coral bottom provides an ideal feeding ground for fish. The fishing boats based at Pensacola and Tampa have frequented this "Middle Ground" for many years to catch red snapper and grouper.

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U. S. COAST GUARD AND GEODETIC SURVEY, WASHINGTON, D.C.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 63 PP. 745-748 JULY 1952

POST-CRETACEOUS FAULTS IN VIRGINIA AND NORTH CAROLINA

BY WILLIAM A. WHITE

It has long been the accepted opinion that late deformation in the Appalachian region has been accomplished wholly by gentle arching, tilting, or uplift, unaccompanied by faulting. Any normal faults in this region have generally been ascribed to the Palisades disturbance, and reverse faults are regularly considered to be at least as old as the Appalachian Revolution. However, the writer has observed several localities in Virginia and North Carolina where faults of small displacement cut young fluvial gravels. Some of these faults are normal; others are reverse.

On U. S. Highway 60, about 1 mile east of its junction with U. S. Highway 220 at Clifton Forge, Virginia, a large road cut about 70 feet high exposes Paleozoic shales. Overlying the bevelled surface of the shales is a remnant of an unconsolidated gravel bed of obviously fluvial origin (Pl. 1, fig. 1) which retains 10 to 12 feet of its original thickness. A nearly vertical fault striking N. 25° W. cuts the shales and the gravel bed. The downthrown block is on the northeast side, and the thickest remnant of the gravel bed forms its surface at the fault. The gravel has been completely eroded from the upthrown block in the immediate vicinity of the fault, but is partially preserved a short distance back from it. A projection of the base of this remnant of the gravel bed suggests that the throw is about 15 to 20 feet. No clearly recognizable horizon in the shale could be identified on opposite sides of the fault.

Clay up to 3–4 inches thick occurs between the surfaces of the fault, both where the two surfaces of the fault were in the shale in the lower part of the exposure, and higher up where the shale of the upthrown block was thrown against the gravel on the downthrown block. Possibly some of this clay was of surface origin, but slickensides were found throughout it.

The slickensided gouge in contact with the gravel bed precludes any possibility that the faulting antedated the deposition of the gravel. Furthermore, since the same type of easily eroded shale both underlies the gravel bed and is faulted against it, on geomorphic grounds alone it would seem highly improbable that it could have localized even a minor fault-line scarp.

The possibility of differential settling is obviated by the fact that no clays or other compressible fluvial materials appear between the gravel and the Paleozoics.

The further possibility that the gravel was deposited contemporaneously with the faulting by deposition in a fault-angle valley is partially negated by the presence of the slickensides, although later movement during the same period of faulting could have produced the slickensides. However, this is a matter of small moment because in either instance the fault must be as young as the gravels.

The age of the gravels is not clear, but they rest on a remnant of what may be the Harrisburg surface at about 1200 feet elevation. They are about 200 feet above the present flood plain of the James River near by, and about 700 feet below the 1900-foot crest level of Wilson Mountain where the James River cuts through the massive quartzites of the Clinch and Clinton formations in Rainbow Gap. Since the gravels are little if any higher than the Harrisburg surface, and are definitely below the general crest levels of the ridges, one would assume they are Tertiary, for any Cretaceous surface that existed here should have been as high as, or higher than, the general ridge crest.

Similar evidence of late high-angle faulting may be seen at the eastern foot of the Blue Ridge near the small community of Hillsboro, on U. S. Highway 250 in Albemarle County, Virginia, about 12 miles west of Charlottesville. In a road cut on the north side of the highway, two minor faults about 20 feet apart cut high-level gravels. The displacement of the more

westerly one is about 3 feet; in the other it is about 1 foot. These gravels are obviously of fluvial origin also, for they rest upon a dissected strath terrace and are overlain by the typical lateral and vertical accretion deposits of flood plains.

Mr. E. O. Gooch called these faults to the writer's attention. They could not be photographed because they were largely grown over by plants set out for road-bank stabilization.

A similar small reverse fault in Stanly County, North Carolina, was called to the writer's attention by Mr. Frank O. Bowman, Jr. It is exposed in a cut along the Norfolk and Southern Railroad about a quarter mile west of the Pee Dee River bridge (Pl. 1, fig. 2). It strikes N. 39° E. and dips about 50° W. Manganese oxide casts of slickensides show motion essentially parallel with the dip. The upthrown block is on the northwest. A bed of fluvial gravel 8 to 10 feet thick rests on Carolina slate and has been displaced 10 to 12 feet by the fault.

These gravels are less than 100 feet above the present level of the Pee Dee River, and it would be difficult to assume that they were older than Pliocene.

The above-mentioned faults show the recency of their movement by cutting remnants of fluvial gravels. Others which cut marine sediments have been observed near the edge of the Coastal Plain. All of them transect small isolated patches of sand or gravel which form outliers of the Coastal Plain where it wedges out over the Piedmont. For this reason it is difficult to determine whether the gravels are remnants of the basal Tuscaloosa or of one of the superficial marine terrace deposits of Pliocene or Pleistocene age.

A small fault (Pl. 2, fig. 1) which cuts such gravels can be seen near the city of Wilson at the edge of the North Carolina Coastal Plain. It was called to the writer's attention by Messrs. Frank O. Bowman, Jr., and James M. Jennings. The writer has not seen this fault, but Messrs. Bowman and Jennings report that it can be seen on the east side of North Carolina Highway 581 about 1 mile north of Buckhorn Crossroads. The strike is N. 51° E.; the dip is 48° W. About 5 feet of coastal-plain sediments overlie Carolina slate. The slate is dragged down on the upthrown block. The displacement seems to have been approximately parallel with the dip and is about 1 foot.

Other late faults which cut coastal-plain gravels are described by Carr (1950) from the northwestern part of the city of Washington, D. C.:

> "A few faults in the bedrock and overlying strata have been noted.... One of the largest, according to N. H. Darton, was revealed about 1925 in a sewer-cut along Eighteenth Street near California Street Northwest. Along this fault, which showed the basal gravels of the Potomac group resting on gneiss, there was a considerable drop on the east side."

Carr does not state whether this is a tensional or thrust fault.

Another small fault is exposed at the intersection of Adams Mill Road and Clydesdale Place near the entrance to the National Zoological Park. This exposure is protected by a screened enclosure and could not be photographed. It seems to be a small thrust striking west of north, and dipping approximately 60° W. About 6 feet of throw is revealed in the exposure, but this may not be complete. The exposure in the apparently upthrown block is wholly of crystalline basement rock, but the apparently downthrown block shows 3–4 feet of gravel resting on crystalline rocks.

Another small fault mentioned by Carr "was formerly exposed in the alley north of Calvert Street about midway between Cliffbourne Place and the Calvert Street Bridge" (Pl. 2, fig. 2).

All the faults described are of minor displacement, but their geomorphic significance may be much greater than their small size suggests. If these faults had not been observed to cut gravels, their late age might never have been

PLATE 1.—FAULTED GRAVELS

FIGURE 1.—FAULT NEAR CLIFTON FORGE, VIRGINIA
Showing fluvial gravels on downthrown block to the right

FIGURE 2.—FAULTED FLUVIAL GRAVELS IN CUT OF NORFOLK AND SOUTHERN RAILROAD
Near Pee Dee River, Stanly County, North Carolina

FIGURE 1

FIGURE 2

FAULTED GRAVELS

FIGURE 1

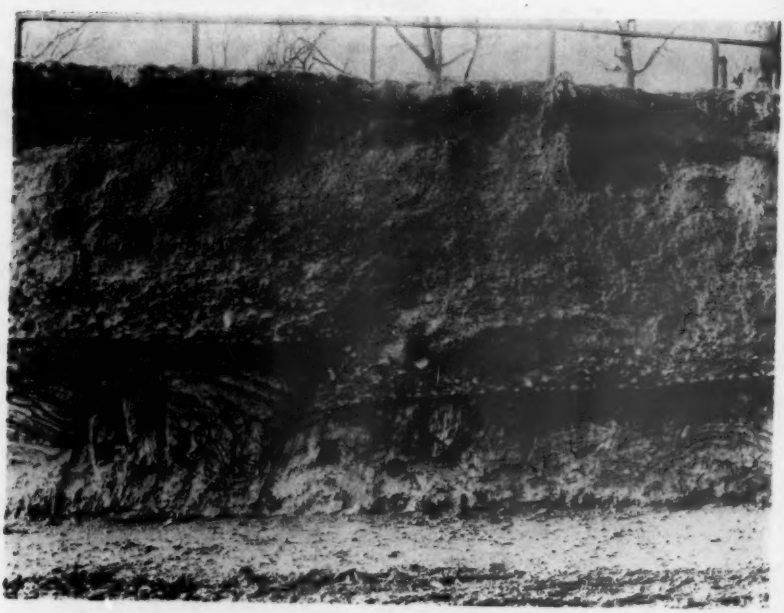

FIGURE 2

FAULTED GRAVELS

suspected, and, if the displaced gravels had not been presented in section by road cuts showing the subjacent rocks, their transection by faults would not have been recognized.

The faults at Clifton Forge and Hillsboro, Virginia, and the one in Stanly County, North Carolina, cut remnants of stream gravels which are preserved only in small isolated patches, and they could readily be components of a large system of late structures which elsewhere do not share their fortuitous location.

The faults in Washington, D. C., and near Wilson, North Carolina, appear at the extreme inner edge of the coastal plain where the gravels are widespread but thin and offer good opportunity for small displacements to throw them visibly against the basement rocks. Also in Washington there has been much excavation which has exposed the faults; and so many geologists live there that even a very briefly exposed fault has an excellent chance of being discovered.

Therefore, these faults may be a few of the minor structures of a system of faulting of considerable magnitude which elsewhere has been confused with earlier faulting such as that of the Triassic.

If the faults in Washington are components of such a larger system, their former topographic manifestation in the coastal plain to the east could readily have been masked by later marine erosion and deposition. McGee's (1888) suggestion of a fault contact between the crystallines and the Cretaceous along the head of Chesapeake Bay is of interest here.

Future geomorphic studies in the Southern Appalachians might consider the influence of such late faulting upon the genesis of topography, and the possibility of its presence in the sediments of the Atlantic Coastal Plain might be considered in the growing search for petroleum there. A number of such faults in the North Carolina Coastal Plain near Cape Hatteras have been suggested by Skeels (1950) from his geophysical data.

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University of North Carolina, Chapel Hill, N. C.

Plate 2.—FAULTED GRAVELS

Figure 1.—Faulted Coastal Plain Gravels Near Wilson, North Carolina

Figure 2.—Faulted Gravels Near Rock Creek Park, Washington, D. C.
After Darton